ON SEISMIC WAVES.

(Third paper.)

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(With 6 figures.)

Contents.

Zusammenfassung: Im Anschluß an Untersuchungen über Oberflächenwellen auf Grund der Seismogramme, die bereits in den beiden vorangehenden Veröffentlichungen [1] [2] benutzt worden waren, wird eine Reihe von allgemeinen Ergebnissen über Oberflächenwellen besprochen. Die Tabellen 1—12 enthalten Messungen der Geschwindigkeiten von Love- und Rayleigh-Wellen. Fig. 2 zeigt das Gesamtergebnis. Die früheren Resultate über den Aufbau der Erdkruste auf Grund solcher Messungen werden im wesentlichen bestätigt und zum Teil verfeinert. So zeigt z. B. Tab. 1, daß deutliche Unterschiede zwischen den verschiedenen tektonischen Einheiten von Nordamerika bestehen. Wahrscheinlich besitzt der Südwesten des Kontinentes eine dünnere kontinentale Kruste als der größte Teil von Nord- und Südamerika. Der Indische und Atlantische Ozean zeigen angenähert gleiche Dispersionskurven;

der Anstieg der Wellengeschwindigkeit mit der Periode der Oberflächenwellen erfolgt schneller als in den Kontinenten, was auf eine dünnere Krustenschicht schließen läßt. Leider sind die Beobachtungen für die Wellengeschwindigkeiten kurzer Oberflächenwellen nicht zahlreich genug, um die Frage entscheiden zu können, ob in diesen Gebieten alle Krustenschichten dünner sind, oder ob die obersten Schichten, besonders die Granitschicht, gänzlich fehlen. Für das eigentliche pazifische Becken ergeben sich bereits für ganz kurze Wellen relativ große Geschwindigkeiten, die keinen Zweifel lassen, daß die oberste Schicht dort aus einem Material besteht, dessen elastische Eigenschaften denen des Sima unter den Kontinentalschollen näherstehen als denen der Kruste. Ob der Boden des pazifischen Beckens stetig in das Sima übergeht, oder ob es sich dort um Schichten mit wenig verschiedenen Eigenschaften handelt, läßt sich nicht entscheiden, doch ist der Unterschied so klein, daß er durch stetige Zunahme der Elastizitätskoeffizienten mit der Tiefe als Folge des zunehmenden Druckes erklärt werden kann. Polynesien und der südwestliche Teil des Pazifischen Ozeans zeigen ähnliche Dispersionskurven wie der Atlantische und Indische Ozean, dagegen entsprechen die Werte für das Nordpolarbecken etwa denen für den Pazifischen Ozean. In der vorangehenden Veröffentlichung [2]) war das gleiche Verhältnis für die Amplituden der reflektierten Longitudinalwellen gefunden worden, so daß anscheinend der Aufbau der beiden zuletzt genannten Gebiete ähnlich ist.

Die Perioden der Nachläufer (Fig. 3—5 und Tab. 13—17) nehmen mit der Entfernung zunächst schneller, später langsamer zu; in den großen Herdentfernungen sowie in den Nachläufern der W_2- und W_3-Wellen herrschen Perioden zwischen 16 und 18 Sekunden überall vor. Die Zunahme der Perioden mit der Entfernung sowie die Absolutwerte hängen, neben anderen Ursachen, vom durchlaufenen Weg ab. Am kleinsten sind die Nachläuferperioden bei Wellen, die nur in Eurasien und im nördlichen Atlantischen Ozean verlaufen.

Besonders eingehend wurden die Amplituden der Oberflächenwellen studiert. In einer Reihe von Beobachtungen wird die schon früher in einzelnen Fällen gefundene Tatsache bestätigt, daß beim Passieren der Umrandung des pazifischen Beckens ein erheblicher Teil der Energie durch Reflexion oder Beugung verlorengeht. Oberflächenwellen, die dieses Grenzgebiet passiert haben, zeigen nur einen Bruchteil der Energie von Wellen, die nur wenig verschieden gelaufen sind, jedoch nicht das pazifische Becken passiert haben. Die gefundenen Extinktionskoeffizienten weichen nur unerheblich von den früher gefundenen Werten ab. Tab. 20 gibt berechnete Amplituden. Dabei ist vorausgesetzt, daß M nicht das pazifische Becken passiert. Bei den langen G-Wellen ist dessen Einfluß nicht erheblich.

Nach Untersuchungen über die Verhältnisse verschiedener Vorläufer und Oberflächenwellen und Unterschiede bei dem gleichen Beben in verschiedenen Azimuten, bedingt durch den Mechanismus bei der Auslösung des Bebens, wird die „Größe“ (magnitude) und Energie der Beben untersucht. Nach der Definition von C. F. Richter ist die „Größe“ M eines Bebens der Briggsche Logarithmus der registrierten maximalen Diagrammamplitude (Mittel der beiden Horizontalkomponenten, ausgedrückt in Mikron) in der Aufzeichnung eines Torsionsseismometers mit den üblichen Konstanten ($T_0 = 0.8$ sec., $V = 2800$, $h = 0.8$) in einer Herdentfernung von 100 km. Die Untersuchungen der vorliegenden Arbeit ermöglichen es, für Beben mit normaler Herdtiefe in einfacher Weise die „Größe" mit relativ hoher Genauigkeit (im allgemeinen auf mindestens $^1/_2$ Einheit genau) anzugeben. Wo Torsionsseismometer vorhanden sind, ist Gleichung (13) zu benutzen, wo A die mittlere Diagrammamplitude (wie oben, aber in mm) ist und $\log A_0$ aus Fig. 6 entnommen wird. Ist die wahre maximale Horizontalbewegung a des Bodens für Oberflächenwellen mit Perioden von 16—20 sec. in Mikron bekannt, so gilt für Herdentfernungen über etwa 25° Gleichung (14). Unter Benutzung dieser Gleichungen wurde die „Größe" einer Reihe von Großbeben berechnet und ein Teil der Ergebnisse in Tab. 24 zusammengestellt. Die Energie E kann dann aus Gleichung (15) berechnet werden. E_0 hat etwa den Wert 10^8 *Erg* mit einer vermutlichen Unsicherheit von einem Faktor 10. Danach hätten Ortsbeben, die gerade noch von einem modernen Instrument mit hoher Empfindlichkeit registriert werden, die Energie 10^8 *Erg* (etwa 1 Meterkilogramm), die größten Beben der Tab. 24 eine Energie von etwa 10^{25} *Erg*.

Zum Schlusse werden die Laufzeiten der seismischen Wogen bei dem Salomon-Inseln-Beben am 3. Oktober untersucht, und es wird ein Überblick über die Resultate betreffend die Schichtung der Erde gegeben.

I. Introductory.

In two previous papers under the present title [1][2] the authors have presented data on bodily waves. The present paper contains the corresponding data on surface waves (including seismic sea waves), the magnitude and energy of earthquakes, and miscellaneous topics. The seismograms used were those studied in the previously published investigations; the numbers assigned to individual shocks are as given in the previous papers.

II. Velocities of surface waves.

On all available seismograms the times of arrival, and the periods of the first surface waves have been measured. These have been classified as Love waves (Q) or Rayleigh waves (R). Theoretically, this classification is possible in two independent ways, based respectively on the horizontal and the vertical components of motion. The horizontal displacement should be transverse for Q waves and longitudinal for R waves, while the vertical component should show only Rayleigh waves. When seismograms for all three components are available at a single station, it is possible to aply both criteria; the results are regularly found to be in excellent agreement. In very numerous instances it is possible to identify both Q and R waves on the same seismograms.

We have used the accepted names of Love and Rayleigh waves for the observed surface waves, although the polarization is not precisely that which this indicates, as small amplitudes are frequently registered in components where they should not be found according to the unmodified theory. This is not surprising, as the theoretically presupposed conditions are never completely realized; for example, the theory requires a homogeneous layering, which is at best a rough approximation to the facts, and occasionally is in serious conflict with them. The surface waves should actually undergo reflection and refraction at the boundaries between tectonic units of crustal structure; and where the path is nearly along such a boundary, exceptional phenomena may occur. Thus certain South American shocks record at Pasadena with two distinct groups of Q waves, the first apparently passing through the Pacific structure, the second through the continent. As the first of these is strictly a refracted wave, the apparent azimuth of displacement on arriving at the station cannot be expected to be exactly transverse to the direction of the epicenter.

It is noteworthy that a considerable number of shocks originating in Polynesia record at Pasadena with definite R but no identifiable Q. This is not invariably the case; especially large shocks, such as the great Solomon Islands earthquake investigated in the present papers, may show a very large long-period Q (G).

It is theoretically possible to draw conclusions regarding the crustal structure from the ratio of the horizontal to the vertical amplitudes in Rayleigh waves. However, no investigation of this kind has been attempted here, as very few cases are available in which seismograms were written in all three components by instruments having approximately the same characteristics, which is necessary for accurate identification and comparison of the three displacements. This difficulty is particularly serious at Pasadena and its auxiliary stations; not only are the characteristics different for the instruments recording the different components, but the constants are in most cases not known with sufficient precision, and in the case of the short-period torsion seismometer there is some doubt whether the principal axes of inertia of the suspended system actually are oriented in the cardinal directions, so that azimuths of displacements determined from these instruments are subject to much uncertainty. Other difficulties in determining the earth motion from the instrumental response are frequent at almost all stations.

Numerical data for the observed velocities and periods are given in Tables 1—6 for Q and in Tables 7—12 for R. The first column of each table gives the serial number of the shock.

Each separately numbered table represents a geographical group, including cases in which the wave paths are principally across the same or similar structural units (continental masses, oceanic basins, etc.). Within the tables a further subdivision is indicated by the horizontal lines.

Separation of this kind is necessary, as the velocities of surface waves necessarily depend on the crustal structure along their paths. The corresponding theory has been extensively discussed by numerous authors. For the shorter periods the velocity of surface waves approaches that of shear waves in the uppermost crustal layer; for longer periods the velocity in general approaches that of shear waves in the layer through which most of the energy is propagated. Consequently, with increasing period and wave length the effect of the deeper layers becomes larger. If, for waves of given period, the velocity of surface waves differs in two given regions, this may be due either to a difference in velocities in corresponding layers, or to different thickness of layers with equal velocities. Analogous considerations apply to the case of gradual increase in velocity with depth.

Theoretically the observed velocities should correspond to the group velocity and not to the phase velocity. For all cases of the type occurring in the problem of seismic surface waves, the group velocity has a minimum for very short periods.

It should be borne in mind that waves of the same character as surface waves may be associated with any of the discontinuities at depth. One case of this sort has recently been investigated by SEZAWA and KANAI[3]). In such instances the wave-length must not be small compared with the thickness of the layer under consideration. If such waves were to occur at the surface of the core, their length consequently must be very great; this is the case for the large U waves found by MACELWANE[4]) in the South Pacific earthquake of 1924, which have been tentatively identified by GUTENBERG[5]) as occurring at the discontinuity between core and mantle.

To dispose of certain theoretical objections[6]) it should be mentioned that in Tables 1—6 the tabulated data for surface shear waves are given only for cases where there is clear evidence of polarization transverse to the direction of propagation.

Table 1.

Velocities v in km./sec. of surface shear waves and residuals d in $^{1}/_{10}$ km./sec. with respect to mean curve for wave paths through America.

Δ = distance in degrees, T = period in seconds.

No.	Epicenter-station	Δ	T	v	d	T	v	d
110	Baja California-Pasadena	7	50±	4.5	4	35	4.1	3
			12	3.6	1			
110	Baja California-Tinemaha	10	30	4.0	3			
111	Gulf of Calif.-Pasadena	$11^1/_2$	30	4.2	5	22	4.1	6
112	Off Mexico-Pasadena	$18^1/_2$	32	4.3	6	22	4.0	5
113	Off Mexico-Pasadena	20	30	3.7	0	17	3.5	0
114	Off Mexico-Pasadena	20	24	3.6	1			
3	Mexico-Pasadena	20	26	4.0	4			
6	Mexico-Pasadena	22	32	4.3	5	25	4.1	5
7	Mexico-Pasadena	$25^1/_2$	40	4.0	1	35	3.9	1
			22	3.6	1			
8	Mexico-Pasadena	$25^1/_2$	40	4.0	1	35	4.0	2
			23	3.8	3			
124	Mexico-Pasadena	20	25	4.0	5			
124	Mexico-Mt. Wilson	20	16	3.7	2			
124	Mexico-Sa. Barbara	$19^1/_2$	17	3.5	0			
124	Mexico-La Jolla	19	20	3.7	2			
126	Mexico-Pasadena	24	40	4.1	2	22	3.8	3
142	Mexico-Pasadena	$17^1/_2$	20	3.7	2			
126	Mexico-Berkeley and Lick	$28^1/_2$	27	3.8	2	22	3.8	3
126	Mexico-Stanford	$28^1/_2$	24	3.8	3			
142	Mexico-Berkeley	23	18	3.3	—2			
1	Texas-Pasadena	12	50	4.1	0	35	3.7	0
			27	3.7	1	18	3.5	0
1	Texas-Haiwee	$12^1/_2$	20	3.6	1	12	3.3	—2
1	Texas-Mt. Wilson	12	17	3.5	0			
10	San Salvador-Pasadena	34	35	3.7	0			
14	Nicaragua-Pasadena	37	42	3.9	—1	36	3.8	0
140	Panama-Pasadena	42	42	4.1	1	38	4.0	1
141	Panama-Pasadena	42	40	3.9	0			
11	E. Alaska-Pasadena	34	42	4.2	2	18	3.7	2
3	Mexico-Sitka	44	44	4.0	0	30	3.8	1
156	Nevada-Sitka	$21^1/_2$	40	4.1	2	32	4.0	3
1	Texas-Sitka	34	23	3.5	0	18	3.4	—1
157	Utah-Sitka	21	28	3.4	—2	12	3.3	—2
154	Alaska-Ottawa	$38^1/_2$	110	4.7	1	60	4.4	0
11	Alaska-Ottawa	44	48	4.0	—1	36	3.7	—1
			10	3.6	1			

Table 1 (continued).

No.	Epicenter-station	Δ	T	v	d	T	v	d
11	Alaska-Halifax	50	50	3.9	—2	40	3.8	—1
			30	3.7	0	10	3.6	1
134	Alaska-Ottawa	44	40	4.0	1	8	3.6	1
134	Alaska-Halifax	$49^1/_2$	9	3.6	1			
134	Alaska-Harvard	$47^1/_2$	55	4.3	1	42	4.0	1
64	Alaska-Technology	$48^1/_2$	48	4.1	0			
156	Nevada-Saskatoon	15	6	3.2	—3			
3	Mexico-Saskatoon	$33^1/_2$	24	3.5	0			
23	New Foundld.-Victoria . .	45	60	4.4	0	14	3.9	4
23	New Foundld.-Pasadena .	48	6	3.6	1			
23	New Foundld.-Haiwee . . .	47	8	3.7	2			
144	Baffin Bay-Pasadena . . .	46	70	4.0	—4	45	3.9	—1
22	Baffin Bay-Pasadena . . .	46	50	4.0	—1	32	3.7	0
			24	3.5	0			
22	Baffin Bay-Tinemaha . . .	43	6	3.6	1			
22	Baffin Bay-Sa. Barbara . .	$46^1/_2$	7	3.8	3			
22	Baffin Bay-Riverside . . .	46	8	3.7	2			
11	Alaska-Charlottesville . . .	48	46	3.9	—1	36	3.7	—1
134	Alaska-Chicago	$40^1/_2$	10	3.6	1			
134	Alaska-Little Rock	44	10	3.5	0			
134	Alaska-Florissant	$41^1/_2$	7	3.6	1			
11	Alaska-Florissant	42	30	3.5	—2	15	3.3	—2
11	Alaska-St. Louis	42	38	3.6	—2	30	3.5	—2
134	Alaska-St. Louis	$41^1/_2$	6	3.8	3			
156	Nevada-Ottawa	32	16	3.6	1			
156	Nevada-Halifax	$40^1/_2$	60	4.5	1	30	3.8	1
158	Utah-Ottawa	$27^1/_2$	16	3.9	4			
157	Utah-Chicago	19	14	3.7	2			
158	Utah-Chicago	19	12	3.6	1			
157	Utah-Little Rock	$17^1/_2$	4	3.5	0			
158	Utah-Little Rock	$17^1/_2$	6	3.7	2	2	3.6	1
158	Utah-Georgetown	$27^1/_2$	13	3.8	3			
1	Texas-Ottawa	$26^1/_2$	7	3.6	1	8	3.5	0
1	Texas-Chicago	17	19	3.5	0	12	3.4	—1
155	Mexico-Ottawa	34	40	3.9	0	30	3.5	—2
6	Mexico-Ottawa	$37^1/_2$	12	3.4	—1			
3	Mexico-Ottawa	$35^1/_2$	8	3.5	0			
6	Mexico-Chicago	$28^1/_2$	16	3.4	—1	13	3.3	—2
124	Mexico-Harvard	$36^1/_2$	12	3.3	—2			
3	Mexico-Georgetown	31	30	3.7	0			
6	Mexico-St. Louis	25	10	3.4	—1			
6	Mexico-Florissant	25	10	3.4	—1			

Table 1 (continued).

No.	Epicenter-station	Δ	T	v	d	T	v	d
3	Mexico-Florissant	23	7	3.3	—2			
3	Mexico-Little Rock	$19^1/_2$	4	3.4	—1			
156	Nevada-San Juan	$49^1/_2$	40	4.1	2	32	3.8	1
3	Mexico-San Juan	36	44	4.1	1	30	3.6	—1
25	Ecuador-Huancayo	11	20	3.2	—2			
24	Colombia-La Paz	25	15	3.9	4			
49	Chile-La Paz	$17^1/_2$	12	3.5	0			
25	Ecuador-La Paz	$18^1/_2$	13	3.5	0			
37	Peru-La Plata	24	42	3.7	—2	32	3.5	—2
49	Chile-La Plata	$12^1/_2$	30	3.6	—1			
25	Ecuador-La Plata	38	22	3.4	—1			
40	Chile-La Plata	18	30	3.9	2	20	3.5	0
40	Chile-Rio de Janeiro	25	10	3.6	1			
49	Chile-Rio de Janeiro	$27^1/_2$	14	3.2	—3			
25	Ecuador-Rio de Janeiro	41	8	3.3	—2			
27	Peru-Rio de Janeiro	$38^1/_2$	60	4.1	—3			
24	Colombia-Rio de Janeiro	$42^1/_2$	8	3.6	1			
24	Colombia-La Plata	$45^1/_2$	25	3.4	—1			

Table 2.

Velocities v in km./sec. of surface shear waves through the basin of the Pacific Ocean, excluding the southwestern part. Δ = distance in degrees. T = period in seconds.

a) From seismograms recorded at Pasadena. * Belongs to the following shock.

No.	Epicenter	Δ	T	v	T	v
13	Hawaii	36	15	4.0		
16	Aleutian Is.	$39^1/_2$	19	4.2	18*	4.4*
115	Aleutian Is.	$40^1/_2$	36	4.4	32*	4.3*
20	Aleutian Is.	43	21	4.3	12	4.1
21	Aleutian Is.	44	17	4.1	16*	4.1*
26	Aleutian Is.	51	27	4.2	23	4.3
30	NW. of Easter I.	58	21	4.4	13	4.2
34	W. of Easter I.	61	21	4.3	8	4.3
36	S. Kamchatka	63	20	4.3		
39	Kurile Is.	68	20	4.3	16	4.3
43	Japan	75	23	4.4	14*	4.3*
44	Japan	75	90	4.6	18	4.4
45	Japan	75	32	4.4	14	4.2
51	Japan	81	25	4.4		

Table 2 (continued).

No.	Epicenter	Δ	T	v	T	v
57	S. of Japan	87	25	4.2		
53	Marianne Is.	84	21	4.4		
55	Marianne Is.	$84^1/_2$	20	4.5		
56	Tonga Is.	87	30	4.4		
59	Solomon Is.	88	25	4.5	22	4.4
60	Solomon Is.	91	35	4.6		
61	Solomon Is.	91	10	4.0		
131	Solomon Is.	$86^1/_2$	25	4.5	14	4.3
132	Solomon Is.	88	18	4.4		
65	Bismarck Is.	93	35	4.3		
123	Bismarck Is.	92	35	4.4		
128	Bismarck Is.	94	40	4.5		
147	Sa. Cruz Is.	$84^1/_2$	40	4.5		
148	Sa. Cruz Is.	85	22	4.3		
66	New Zealand	$93^1/_2$	40	4.5		
67	New Zealand	$94^1/_2$	30	4.4		
119	New Zealand	96	20	4.3		
79	Philippine Is.	$104^1/_2$	45	4.4	35	4.5
84	Philippine Is.	$107^1/_2$	30	4.5		
133	Philippine Is.	106	35	4.5	25	4.4

b) From seismograms recorded at Huancayo, Peru.

No.	Epicenter	Δ	T	v	T	v
15	NW. of Easter Is.	38	16	4.4		
183	W. of Galapagos Is.	$31^1/_2$	10	4.3		
160	Near Samoa	93	40	4.5		
118	New Hebrides	112	24	4.3		
147	Sa. Cruz Is.	114	65	4.4		
131	Solomon Is.	116	40	4.3	28	4.3
132	Solomon Is.	119	48	4.4		
123	Bismarck Is.	$129^1/_2$	48	4.4		
173	Bismarck Is.	134	60	4.3	32	4.5
181	Bismarck Is.	130	70	4.4		
128	Bismarck Is.	130	60	4.4		
98	Off New Guinea	$135^1/_2$	48	4.5		
133	Philippine Is.	157	55	4.5	70	4.8 ?

Table 2 (continued).

c) From seismograms recorded at various stations.

No.	Epicenter-station	Δ	T	v	T	v
44	Japan-Tinemaha	72	40	4.5	23	4.4
131	Solomon Is.-Sitka . . .	85	16	4.4		
13	Hawaii-Victoria	38	15	4.2		
34	Easter Is.-Victoria . . .	75	40	4.3		
169	Mexico-Honolulu	58	12	4.2		
170	Mexico-Honolulu	50	24	4.6		
171	Mexico-Honolulu	50	8	4.0		
112	Mexico-Honolulu	$46^1/_2$	25	4.5	15	4.3
172	Mexico-Honolulu	50	15	4.4		
49	Chile-Wellington	83	30	4.4		
26	Aleutian Is.-La Jolla . .	53	31	4.3		
44	Japan-Sa. Barbara . . .	74	18	4.4		

d) From seismograms of the Solomon Is. shocks Oct. 3, 1931. Data from the main aftershock are marked by*.

Station	Δ	T	v	T	v
Honolulu	51	11	4.0	24*	4.5*
Hawai Volcano Obs.	52	36	4.5	33	4.2
Sitka	85	60	4.5	16	4.0
		30*	4.5*		
Ukiah	$85^1/_2$	28*	4.5*		
Stanford	$85^1/_2$	30	4.4	18	4.3
Berkeley	$85^1/_2$	45	4.5	19	4.4
		32*	4.1*		
Santa Clara	86	38	4.1		
Lick Obs.	86	45	4.5	20	4.4
		20*	4.4*		
Santa Barbara	$86^1/_2$	54	4.5		
Pasadena.	88	46	4.5	17*	4.4*
Mt. Wilson	88	50	4.5	12	4.2
Victoria	88	53	4.5	45	4.5
La Jolla	$88^1/_2$	48	4.5		
Riverside.	$88^1/_2$	50	4.5		
Haiwee.	$88^1/_2$	52	4.5		
Seattle.	$88^1/_2$	52	4.5		
Tinemaha	$88^1/_2$	48	4.5		
Santiago de Chile	114	70	4.5		
La Paz	124	100	4.6	60	4.5

Table 3.

Velocities v in km./sec. of surface shear waves, originating from the Solomon Is. shocks, October 3, 1931. Polynesian paths. Data from the main aftershock are marked by*.
Δ = distance in degrees. T = period in seconds.

Station	Δ	T	v	T	v
Phu Lien	$62^1/_2$	41	4.2	25*	3.7*
Hongkong	57	37	4.2	26*	4.3*
Zikawei	$56^1/_2$	55	4.4	22*	3.9*
Kobe	$51^1/_2$	54	4.3	47	4.3
Tokyo	$50^1/_2$	54	4.1	45	4.0
		23	3.8		
Manila	$47^1/_2$	19	3.5		
Apia	26	12	4.1	15*	4.0*

Table 4. Velocities v in km./sec. of surface shear waves with Atlantic paths.
Δ = distance in degrees. T = period in second.

No.	Epicenter-station	Δ	T	v	T	v
32	North Atlantic-Scoresby Sund	$15^1/_2$	11	3.9	9	3.7
32	North Atlantic-Ivigtut	$8^1/_2$	12	3.8		
46	East Atlantic-Scoresby Sund .	$31^1/_2$	22	4.0		
42	Central Atlantic-Scoresby Sund	60	32	4.3	19	4.2
35	Central Atlantic-Technology .	$66^1/_2$	22	4.4		
76	Central Atlantic-Harvard . . .	$69^1/_2$	28	4.1		
35	Central Atlantic-San Juan . .	56	32	4.1	20	4.0
46	East Atlantic-San Juan . . .	47	40	4.5	32	4.3
96	South Atlantic-Scoresby Sund	124	30	4.3!		
94	South Atlantic-Rio de Janeiro	$38^1/_4$	28	4.1		
95	South Atlantic-Rio de Janeiro	43	18	4.1		
96	South Atlantic-Rio de Janeiro	$41^1/_2$	20	4.0		
94	South Atlantic-La Plata . . .	33	40	4.5	25	4.2
95	South Atlantic-La Plata . . .	38	22	4.2		
96	South Atlantic-La Plata . . .	$39^1/_2$	22	4.4		

Table 5.

Velocities v in km./sec. of surface shear waves, paths through the Indian Ocean. Epicenters of shocks No. 107 and 108 are in the Indian Ocean, southeast of Madagascar.
An * in the first column indicates data from the Sumatra shocks, Feb. 10 or Sept. 25, 1931.

No.	Station	Δ	T	v	T	v
107	Wellington	86	50	4.4	30	4.3
107	Melbourne	$69^1/_2$	45	4.4		

Table 5 (continued).

No.	Station	Δ	T	v	T	v
106, 107	Perth	49	11	4.2	14	4.0
107	Colombo	46	16	3.9		
107, 108	Bombay	$53^1/_2$	20	4.1	20	4.0
107	Adelaide	65	33	4.0		
*	Tananarive	$55^1/_2$	29	4.5	25	4.4
*	Johannesburg	74	45	4.1	25	4.0
*	Johannesburg	74	33	4.2		

Table 6.

Velocities v in km./sec. of surface shear waves with paths partly across the north polar region.

Epicenters in Alaska.

Epicenter-station	No. 134			No. 136			No. 138		
	Δ	T	v	Δ	T	v	Δ	T	v
Abisko . . .	51	36	4.4	56	26	4.0	53	36	4.1
Upsala . . .	58	45	4.2	$64^1/_2$	40	4.1	62	36	4.2
Helsingfors .	$57^1/_2$	40	4.3	$64^1/_2$	24	4.3			

Table 7.

Velocities v in km./sec. of Rayleigh waves through America.

Δ = distance in degrees. T = period in seconds.

No.	Epicenter-station	Δ	T	v	T	v
154	Mexico-Pasadena	$18^1/_2$	30	3.7	22	3.6
153	Honduras-Pasadena . . .	34	40	3.8		
14	Nicaragua-Pasadena . .	37	30	3.4	25	3.4
124	Mexico-La Jolla	19	17	3.4		
142	Mexico-Pasadena	$17^1/_2$	25	3.6	16	3.3
140	Panama-Pasadena . . .	42	25	3.4		
141	Panama-Pasadena . . .	42	30	3.4	25	3.4
144	Baffin Bay-Pasadena . .	46	70	4.0	45	3.9
			22	3.5		
155	Mexico-Ottawa	34	45	3.8		
1	Texas-Charlottesville . .	22	24	3.4		
158	Utah-Florissant	18	12	3.4	7	3.1
6	Mexico-Florissant . . .	25	30	3.4		
6	Mexico-St. Louis	25	12	3.1		
158	Utah-St. Louis	18	10	3.1		
134	Alaska-St. Louis	42	17	2.9		
158	Utah-Georgetown . . .	$27^1/_2$	8	3.2		
3	Mexico-Georgetown . . .	31	24	3.3		
124	Mexico-Harvard	$36^1/_2$	8	2.9		

Table 7 (continued).

No.	Epicenter-station	Δ	T	v	T	v
25	Ecuador-Huancayo . .	11	12	2.8		
24	Colombia-La Paz. . . .	25	15	3.0		
37	Peru-La Plata	24	15	3.1		
49	Chile-La Plata.	$12^1/_2$	30	3.2		
25	Ecuador-La Plata . . .	38	30	2.7		
40	Chile-La Plata.	18	18	2.8		
24	Colombia-Rio de Janeiro	$42^1/_2$	10	3.1		

Table 8.

Velocities v in km./sec. of Rayleigh waves through the basin of the Pacific Ocean, excluding the southwestern part.

a) From seismograms recorded at Pasadena.

No.	Epicenter	Δ	T	v	T	v
13	Hawaii	36	21	4.0		
115	Aleutian Is.	$40^1/_2$	25	4.0	17	3.9
21	Aleutian Is.	44	28	4.0		
29	Kamchatka	57	30	4.0	23	3.7
36	S. Kamchatka.	63	20	3.8		
116	Near Samoa.	71	30	4.0	21	3.9
117	Near Samoa.	71	34	4.0	24	3.9
127	SW. of Samoa	72	25	4.1		
43	Japan	75	25	4.0		
150	S. of Japan.	89	36	4.1	32	4.1
129	Marianne Is..	$84^1/_2$	40	4.1	30	4.1
54	Kermadec Is.	$84^1/_2$	33	4.0	20	3.7
118	New Hebrides	83	32	4.0		
121	New Hebrides	87	25	4.0		
130	New Hebrides	86	24	4.0		
59	Solomon Is..	88	40	4.1	30	4.0
60	Solomon Is..	91	33	4.2	22	4.1
61	Solomon Is..	91	25	3.9		
62	Solomon Is..	$91^1/_2$	35	4.2		
63	Solomon Is..	$91^1/_2$	35	4.0		
131	Solomon Is..	$86^1/_2$	30	4.1	24	4.0
132	Solomon Is..	88	33	4.1	27	4.0
147	Sa. Cruz Is..	$84^1/_2$	35	4.1		
148	Sa. Cruz Is..	85	36	4.0		
149	Sa. Cruz Is..	$84^1/_2$	35	4.2		
183	W. of Galapagos Is. . .	$41^1/_2$	22	3.9		
166	S. Pacific	87	35	4.1		
182	S. Pacific	73	26	4.0		
65	Bismarck Is.	93	35	4.0	26	3.9

Table 8 (continued).

No.	Epicenter	Δ	T	v	T	v
123	Bismarck Is.	92	34	4.1	22	3.9
128	Bismarck Is.	94	30	4.0		
135	Bismarck Is.	92	29	4.0		
145	Bismarck Is.	95	35	4.0		
66	New Zealand	$93^1/_2$	25	4.0		
119	New Zealand	96	30	4.1	28	4.0
74	New Guinea	$100^1/_2$	30	4.1	24	3.8
77	Philippine Is.	104	30	4.0		
79	Philippine Is.	$104^1/_2$	30	3.9	25	3.8
83	Philippine Is.	$107^1/_2$	35	4.1		
84	Philippine Is.	$107^1/_2$	35	4.1		
133	Philippine Is.	106	30	4.0		
85	Off Celebes	111	68	4.0	40	4.0

b) From seismograms recorded at Huancayo, Peru.

No.	Epicenter	Δ	T	v	T	v
15	NW. of Easter I.	38	12	3.9		
182	SE. Pacific	$32^1/_2$	25	4.0		
166	S. Pacific	52	30	4.1	24	4.0
119	New Zealand	$94^1/_2$	30	4.0		
179	New Zealand	95	30	4.0		
184	New Zealand	95	28	4.0		
127	SW. of Samoa	96	24	4.0		
118	New Hebrides	112	30	3.8		
121	New Hebrides	112	30	4.0		
131	Solomon Is.	116	30	3.9		
132	Solomon Is.	119	30	4.0		
98	New Guinea	$135^1/_2$	36	4.0		
100	Japan	137	33	3.9		
129	Marianne Is.	139	48	4.0	28	3.9
162	Bonin Is.	140	32	4.0		
133	Philippine Is.	157	48	4.0	30	3.9

c) From seismograms recorded at various stations.

No.	Epicenter-station	Δ	T	v	T	v
26	Aleutian Is.-La Jolla	53	25	4.0		
116	Samoa-Berkeley	69	30	4.1		
116	Samoa-Victoria	74½	30	4.0		
169	Mexico-Honolulu	58	30	4.1		
171	Mexico-Honolulu	50	25	4.0		
34	W. of Easter I.-Berkeley	64	27	3.9		

Table 8 (continued).

No.	Epicenter-station	Δ	T	v	T	v
40	Chile-Wellington	$93^1/_2$	30	4.0	24	3.9
48	Chile-Wellington	$85^1/_2$	35	4.1		
49	Chile-Wellington	83	30	4.0		
40	Chile-Christchurch . . .	$93^1/_2$	30	4.0		

d) From seismograms of the Solomon Island shocks, October 3, 1931. Data from the main aftershock are marked by*.

Station	Δ	T	v	T	v
Berkeley.	$85^1/_2$	34	4.1		
Santa Barbara	$86^1/_2$	33	4.1	32*	4.1*
Pasadena	88	35	4.0	28	4.0
		25*	3.9*		
La Jolla.	$88^1/_2$	30	4.0		
Tinemaha	$88^1/_2$	35	4.0		

Table 9.

Velocities of Rayleigh waves originating from the Solomon Island shocks, October 3, 1931; Polynesian paths.

Data from the main aftershock are marked by*.

Station	Δ	T	v	T	v
Kobe	$51^1/_2$	19	3.5		
Tokyo.	$50^1/_2$	50	3.8	18	3.5
Apia	26	22	3.7	7*	3.6*
Riverview	25	33	3.9		

Table 10.

Velocities v in km./sec. of Rayleigh waves with Atlantic paths.

No.	Epicenter-station	Δ	T	v	T	v
96	South Atlantic-Scoresby Sund	124	22	3.7		
32	North Atlantic-Scoresby Sund	$15^1/_2$	8	3.3		
32	North Atlantic-Ivigtut .	$8^1/_2$	20	3.7	12	3.4
32	North Atlantic-Harvard	27	22	3.4		
94	South Atlantic-La Plata	33	30	3.9		

Table 11.
Velocities v in km./sec. of Rayleigh waves with paths through the Indian Ocean.
Same epicenters as in table 5.

No.	Station	Δ	T	v	T	v
106, 107	Wellington	86	27	3.9	30	4.0
107	Melbourne	$69^1/_2$	24	3.9		
107	Christchurch	83	30	4.0	27	3.9
106, 107	Batavia	53	23	3.9	20	4.1
107, 108	Colaba-Bombay . . .	$53^1/_2$	22	3.9	22	3.8
Sumatra	Tananarive	$55^1/_2$	27	4.0	27	4.0

Table 12.
Velocities v in km./sec. of Rayleigh waves with paths partly across the north polar region.
Epicenters in Alaska.

Epicenter-station	No. 134			No. 136		
	Δ	T	v	Δ	T	v
Scoresby Sund	45	30	3.5			
Abisko	51	30	3.7	56	24	3.7
Upsala	58	30	3.8	$64^1/_2$	28	3.6

Table 1 gives the data for Q waves across the American continent. In Fig. 1 these data are plotted as diagonal crosses (North America) and triangles (South America). In the figure a representative curve has been drawn through these points; the residuals of the individual readings with respect to this curve are given in the columns headed d in Table 1! The beginning of this curve is doubtful. Theoretically, the velocity should approach 3.2 km./sec. for shorter periods in regions such as Southern California[7] [8] [9], where the granitic rocks either are at the surface, or else form a basement underlying a thin layer of other rocks, so that the prevailing velocity of transverse waves in the uppermost layers is the 3.2 characteristic of granite. Even in such regions, Q waves of higher velocity may be observed as a result of propagation along a discontinuity at depth. In eastern North America, velocities of 3.5 and 3.6 km./sec. have been observed at the surface[10] [11] [12]; if this proves to be the prevailing velocity near the surface over a large region, then lower velocities for Q should not be observed at all in that area.

The first section of Table 1, above the first horizontal dividing line, contains data referring to wave-paths which lie chiefly in the

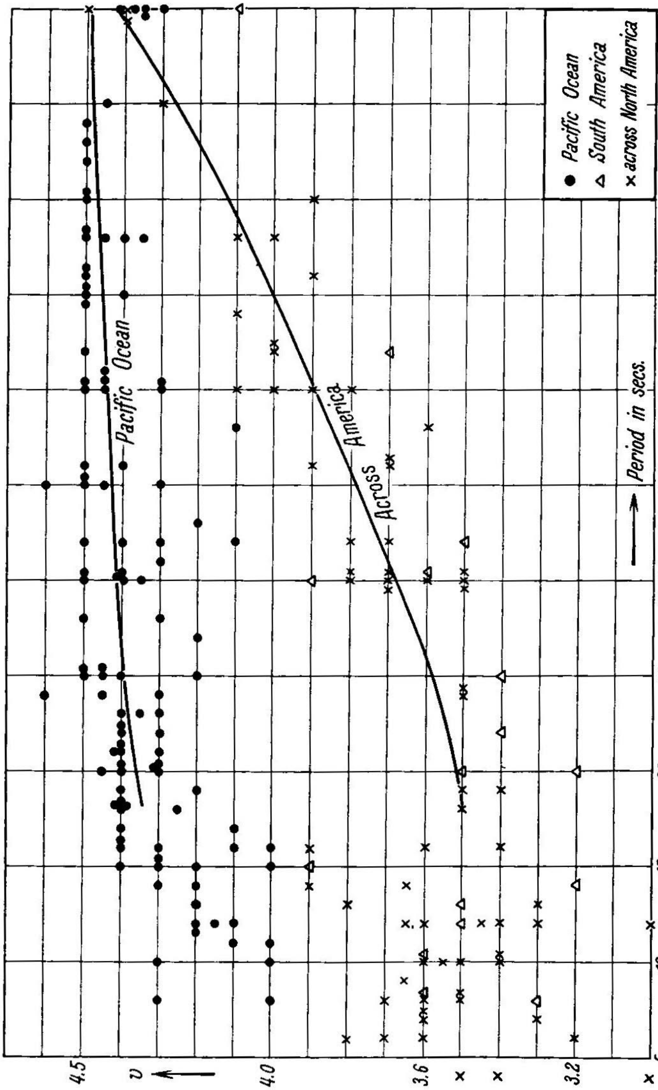

Fig. 1. Velocity of Love waves (Q) as a function of period.

western coastal region of North America. There is a marked prevalence of positive residuals with respect to the mean curve, which does not occur for any of the other geographical groups of Table 1. Only seismograms from the shorter distances are included in this section, since the particular structure characteristic of the western coastal region extends only to a limited distance in any one azimuth. Consequently, there are no data on the shorter periods in surface waves across this region, since such short-period waves cannot be distinguished from the S waves. Accordingly it is not possible to bring the data on surface waves into relation with the observed velocity of 3.2 km./sec. for bodily waves in the upper layer of this region. The fact that the data for longer periods indicate higher velocities than the mean for the continent may indicate either that higher velocities occur at moderate depth, or more probably that the low-velocity layers are thinner than elsewhere.

The following three groups in Table 1 refer to paths only slightly farther inland than those of the first group; but the observed velocities show only slight deviations from the mean, indicating that the condition which produces the higher velocities of the first group is found only in the western coastal region. The remaining data for paths across North America show no large residuals. The data for the Nevada and Utah shocks at eastern stations appear to indicate slightly high velocities.

In South America the data appear to indicate velocities for Q waves slightly smaller than those for North America.

Table 2 gives the observed velocities for Q waves with paths across the basin of the Pacific Ocean (excluding the Polynesian area); these are plotted as solid circles in Fig. 1. As appears from the figure, these observations show very little scatter, except for the very shortest periods. The data are in good agreement with those previously found by several investigators[13]). The very plainly indicated difference between the observed velocities for oceanic and continental paths leaves no doubt that these two units have entirely different surface structures. The observed curve for Pacific paths can be explained most probably, as has been usual[14]), by supposing that the crustal layers characteristic of the continents are absent from the Pacific basin, and that the velocity increases gradually with depth, which is to be expected. An alternative explanation, as suggested by Byerly[15]), assumes a surface layer in the oceanic area with a velocity for S waves of 4 km./sec., which is not very different from the corresponding velocity for the deepest of the continental crustal layers.

Table 3 contains a few observed velocities for Q waves over Polynesian paths. They are clearly very much lower than the velocities across the Pacific basin proper, and are only slightly higher than those found on the west coast of North America. These data leave no doubt that the structure of the Polynesian area differs considerably from that of the oceanic basin to the north and east of it, and that its uppermost rocks have elastic properties close to those of rocks occurring in the continental crustal layers. The term "Polynesian" has been used in a somewhat extended sense, as only the observation at Apia refers to a path through the eastern island region. The other paths chiefly lie through the triangular area west of the Marianne Islands.

Tables 4 and 5 show the velocities of surface shear waves across the Atlantic and Indian Oceans. These velocities are intermediate between those for the continents and the Pacific Ocean. There are no data for the shorter periods, so that no conclusions are possible as to the upper crustal layers; but in both oceanic areas there must exist crustal layers with properties approaching those of continental layers. With increasing periods the velocities in these cases approach those for the Pacific region more rapidly than is the case for the corresponding velocities across the continents; this indicates that crustal layers underlying the Indian and Atlantic Oceans are thinner than the continental crust.

Table 6 gives the corresponding data for the north polar region. These few observations are similar to those of Tables 4 and 5; but it must be considered that the wave paths now include considerable continental segments both near the epicenters and near the stations. It follows that the velocities through the polar basin itself must be close to the velocities across the Pacific basin. This agrees with the results found in the preceding paper from observations of the amplitude ratio PP/P[2]), and makes it very probable that there is no true continental crust in a large part of the north polar basin.

The velocities for Q waves are plotted in Fig. 2a, with separate curves for each region, with the addition of a curve for Eurasia taken from previous observations.

Observations of Rayleigh waves are given in a similar way as was done for Q waves; representative curves are plotted in Fig. 2b. They are similar to those of Fig. 2a.

III. Periods of surface waves.

In normal seismograms the surface waves begin with the long-period waves which have been discussed in the preceding section. These are followed by waves of shorter and shorter period, the registered amplitudes gradually increasing to the maximum of the seismogram. On computation it is frequently found that the true maximum of

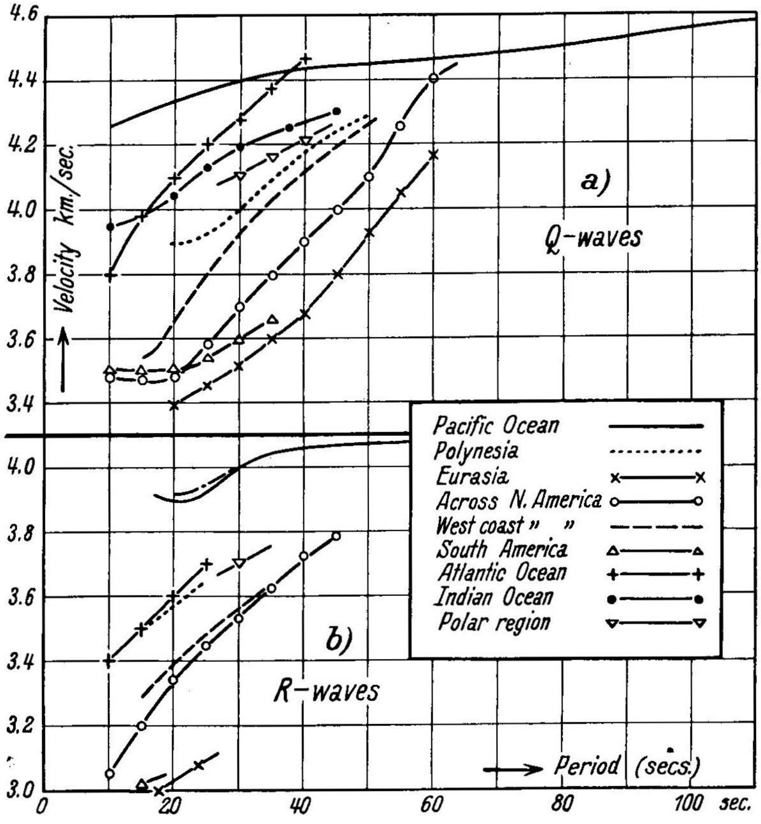

Fig. 2. Velocity of Love waves (Q) and Rayleigh waves (R) as a function of period.

ground amplitude already occurs in the initial long waves (G, R), and that the apparent maximum is in large part due to higher instrumental magnifications for the shorter periods. Following the registered maximum the motion usually becomes more regularly sinusoidal, and continues as the cauda of the seismogram, in which definite prevailing periods are usually distinguishable. Especially at the larger distances, these waves are superposed on the groups of surface waves returning over the major arc, etc. (W_2 etc.) with their respective trains of caudal waves.

For the first ten degrees the periods of the maxima increase considerably with distance[16]; beyond this the increase is more gradual[17] [18], and at about 90° the period of the maximum group becomes constant and remains so into the range of the W_2 and W_3 waves. The periods of these waves in the Sumatra and Solomon Islands earthquakes were from 16 to 22 seconds. At the larger distances no maxima with periods appreciably shorter than these have been observed in these shocks. On the other hand, the longer-period waves become increasingly prominent at very large distances, as their amplitudes fall off less rapidly with distance than is the case for waves of shorter period.

The prevailing periods in the cauda have been measured for all available seismograms. For shocks recorded at Pasadena these are given in Table 13; for shocks recorded at Huancayo, in Table 14; for selected shocks at various stations, in Table 15; for the Sumatra shocks, in Table 16; and for the Solomon Islands shocks, in Table 17. The data of the last two tables are mapped as Figs. 3 and 4. In regions where there are a large number of observations, as in Europe, only a few representative values have been entered on these maps. In both cases there is a clear increase in period with distance, but no evident effect of azimuth.

The effect of distance is also shown in Fig. 5, which in addition to the data of the present paper includes curves based on the results of previous investigations[19] [20] [21]. There are three separable groups of curves, indicated in the figure as I, IIa, and IIb. Group I includes data for shocks recorded at Pasadena from epicenters lying to the northeast, and also for two Japanese earthquakes as recorded in Europe. This group shows decidedly smaller periods in the cauda than any of the others. Group IIa includes data on shocks reaching Pasadena over Pacific paths, as well as most of the shocks both originating and recorded in North America (first section of Table 15); and, in separate curves, the average data from the Sumatra shocks and (for large distances only) the Atacama (Chile) shock as recorded in Europe. The remaining data, represented by Group IIb, are chiefly from Pacific paths.

It is clear that with increasing distance the prevailing periods in the cauda increase, and that the increase is more gradual at the larger distances. The cauda of the W_2 or W_3 waves usually shows periods of about 17 or 18 seconds. On the other hand, there is a clear effect of path, the nature of which is not at first evident. It is obvious that

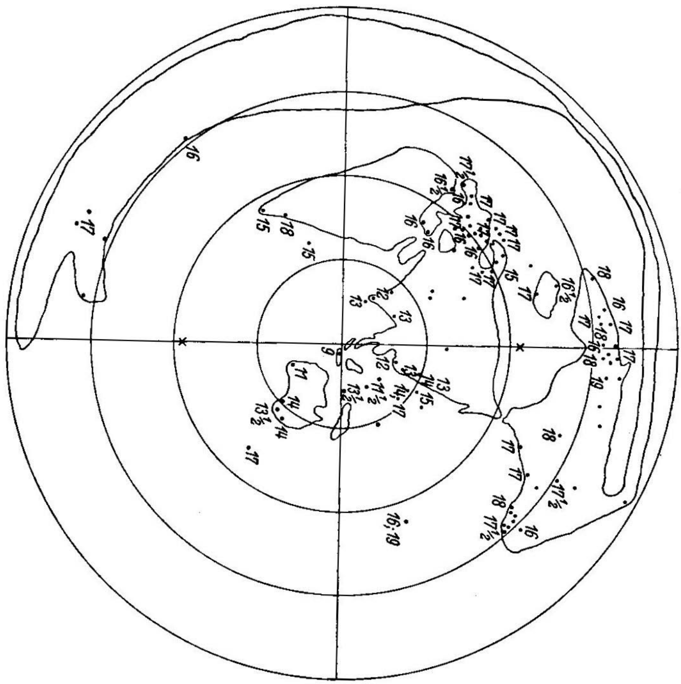

Fig. 3. Periods of cauda waves in the Sumatra shock, September 25, 1931.

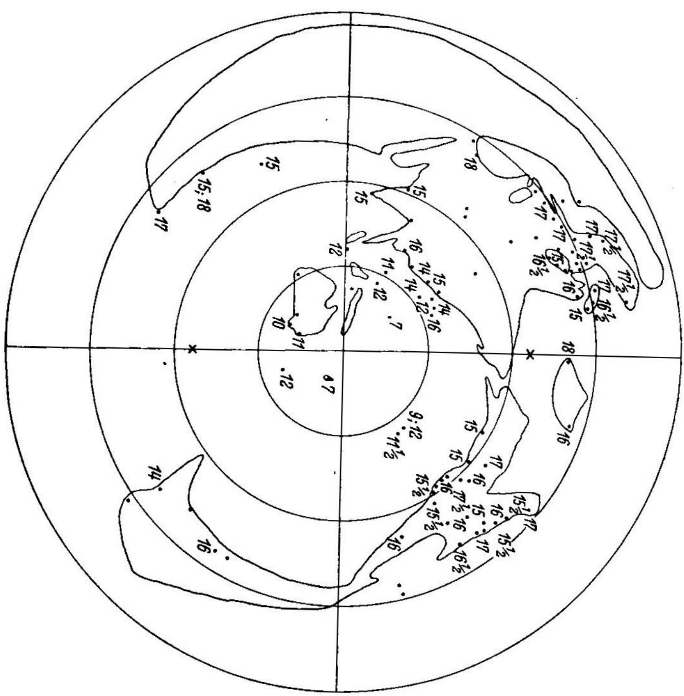

Fig. 4. Periods of cauda waves in the Solomon Islands earthquake of October 3, 1931.

Table 13.
Prevailing periods (seconds) in the cauda of earthquakes recorded at Pasadena.

No.	Epicenter	Dist. degr.	Period	No.	Epicenter	Dist. degr.	Period
109	Gulf of Calif. . .	5	5—8	48	Chile	79	14
110	Baja Calif. . . .	7	6—8	49	Chile	80	$13^1/_2$
111	Gulf of Calif. . .	11	8	50	Chile	80	16
1	Texas.	12	7	51	Japan	81	10;15
112	Off Mexico . . .	19	8	118	New. Hebrides .	84	16
113	Off Mexico . . .	20	8	52	N. Marianne Is. .	84	$17^1/_2$
114	Off Mexico . . .	20	7	53	Marianne Is. . .	84	16
3	Mexico	20	9	54	Kermadec. . . .	84	17
7	Off Mexico . . .	26	8	55	Near Guam . . .	85	16
8	Off Mexico . . .	26	11	56	Tonga Is.	87	16
10	San Salvador . .	34	10	57	S. Japan	87	17
11	Alaska	34	10	58	Solomon Is. . . .	88	$16^1/_2$
12	Alaska	35	12	59	Solomon Is. . . .	88	16
13	Hawaii	36	7; 12	60	Solomon Is. . . .	91	17
14	Nicaragua. . . .	37	13	62	Solomon Is. . . .	91	17
16	Aleutian Is. . . .	40	11	63	Solomon Is. . . .	92	17
17	Cuba	40	10	65	Bismarck Is. . .	93	17
115	Aleutian Is. . . .	41	11;13	66	New Zealand . .	94	16
20	Aleutian Is. . . .	43	13	67	New Zealand . .	95	16
21	Aleutian Is. . . .	44	9	119	New Zealand . .	96	$16^1/_2$
22	Baffin Bay . . .	46	9	70	Altai	95	15
23	New Foundld. . .	48	14	71	Altai	96	$14^1/_2$
24	Colombia	49	13	72	Greece	98	$14^1/_2$
25	Ecuador.	51	$11^1/_2$	120	Kansu	99	15
26	Aleutian Is. . . .	51	13	73	Kansu	100	14
27	Peru	54	12	74	New Guinea. . .	101	16
28	Peru	$54^1/_2$	13	75	Kos	102	15
30	Near Easter I.. .	58	12	122	SW. Atlantic . .	102	13
34	W. of Easter I. .	61	7	80	Szechwan	103	16
36	Kurile Is.	63	10	77	Philippine Is. . .	104	17
39	Kurile Is.	68	10	79	Philippine Is. . .	105	17
40	Chile	70	12	133	Philippine Is. . .	106	17
116	Near Samoa. . .	71	$15^1/_2$	84	Philippine Is. . .	108	$17^1/_2$
117	Near Samoa. . .	71	$15^1/_2$	85	Off Celebes . . .	111	$16^1/_2$
41	Chile	72	14	87	Banda Sea . . .	113	$16^1/_2$
42	Centr. Atlantic. .	74	15	88	Celebes Sea . . .	114	17
43	Japan	75	15	83	Philippine Is. . .	108	$16^1/_2$
44	Japan	75	$9^1/_2$	89	India	115	17
45	Japan	75	9;18	91	Baluchistan . . .	116	$14^1/_2$
46	E. Atlantic . . .	78	14	92	Baluchistan . . .	116	16
47	Chile	78	$13^1/_2$	146	Baluchistan.. . .	117	$16^1/_2$

Table 13 (continued).

No.	Epicenter	Dist. degr.	Period	No.	Epicenter	Dist. degr.	Period
94	S. Atlantic . . .	120	15	130	New Hebrides . .	86	17
97	Sumatra	131	18	150	S. of Japan . . .	89	17
101	Off Ceylon . . .	140	18	124	Mexico	20	8
107	W. Indian Oc.. .	176	17	142	Mexico	18	9
108	W. Indian Oc.. .	176	$15^1/_2$	154	Mexico	19	$7^1/_2$
121	New Hebrides . .	87	$15^1/_2$	153	Honduras	34	10
123	Bismarck Is. . .	92	$16^1/_2$	126	S. Mexico	24	$9^1/_2$
128	Bismarck Is. . .	94	$16^1/_2$	140	Panama.	42	12
135	Bismarck Is. . .	92	17	141	Panama.	42	12
145	Bismarck Is. . .	95	17	136	Alaska	34	13
147	Sa. Cruz Is.. . .	85	16	138	Alaska	33	11
148	Sa. Cruz Is.. . .	85	$16^1/_2$	151	Alaska	36	11
149	Sa. Cruz Is.. . .	85	16	143	Iceland	63	11
131	Solomon Is.. . .	87	16	125	Baffin Bay . . .	46	8
132	Solomon Is.. . .	88	16	144	Baffin Bay . . .	46	$9^1/_2$
127	SW. of Samoa. .	72	16	134	Alaska	34	8;13
129	Marianne Is.. . .	85	17				

Table 14. Prevailing periods in the cauda of earthquakes recorded at Huancayo.

No.	Epicenter	Dist. degr.	Period sec.	No.	Epicenter	Dist. degr.	Period sec.
40	Chile	10	$9^1/_2$	134	Alaska	92	17
25	Ecuador.	11	10	64	Alaska	93	15
2	Chile	20	11	160	Near Samoa. . .	93	15
4	Galapagos Is. . .	21	7	119	New Zealand . .	95	$15^1/_2$
9	Nicaragua	26	$12^1/_2$	164	Off Zululand . .	100	16
183	Galapagos Is. . .	31	7	115	Aleutian Is. . . .	102	18
159	Mexico	39	11	78	Greece	105	16
124	Mexico	40	14	121	New Hebrides . .	112	15
154	Mexico	43	12	107	SW. Indian Oc. .	116	18
19	Mexico	42	$12^1/_2$	131	Solomon Is. . . .	116	16
15	Near Easter I.. .	38	11	132	Solomon Is. . . .	119	16
122	SW. Atlantic . .	45	$13^1/_2$	93	Kamchatka . . .	119	18
167	S. Atlantic . . .	58	15	123	Bismarck Is. . .	130	16
175	Lower California .	58	$13^1/_2$	98	Off New Guinea .	136	$16^1/_2$
176	Lower California .	58	14	161	Off N. Japan . .	136	18
94	S. Atlantic . . .	60	15	162	Bonin Is.	140	19
35	Off Africa . . .	61	$12^1/_2$	102	Kansu	151	17
156	Nevada	65	$14^1/_2$	103	Philippine Is. . .	155	19
38	S. Atlantic . . .	66	16				

there are cases in which the periods of the cauda are very different when the crustal structures traversed are quite similar, and other

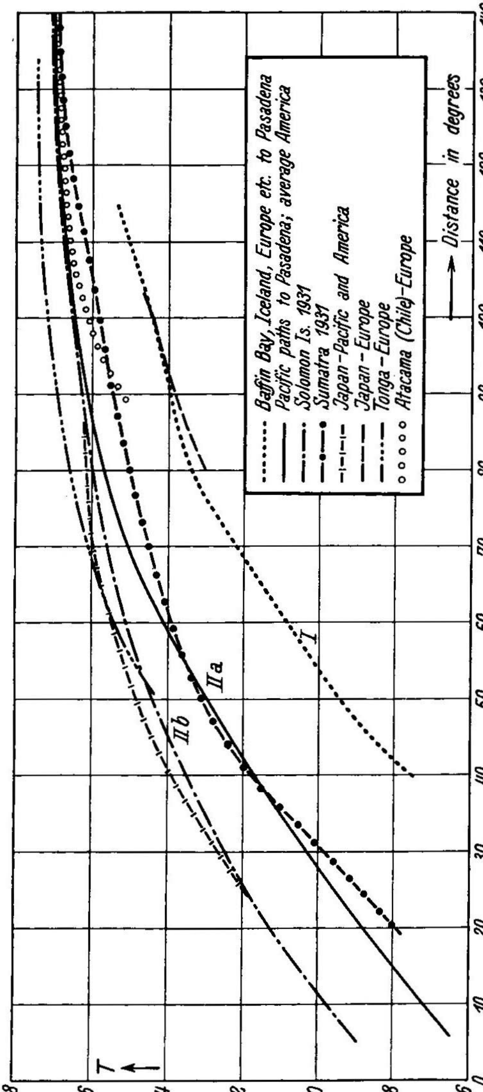

Fig. 5. Prevailing periods T of the cauda as a function of the distance.

cases in which there is a marked similarity in the periods when entirely different structures have been traversed.

It has been suggested [21] [22] that the prevailing period in the cauda is related to the period for which the group velocity has a min-

Table 15.

Prevailing periods T in seconds in the cauda of selected seismograms.

No.	Epicenter-station	Dist. degr.	T	No.	Epicenter-station	Dist. degr.	T
154	Alaska-Ottawa	39	11	3	Mexico-Ottawa	36	11
11	Alaska-Ottawa	44	$11^1/_2$	155	Mexico-Ottawa	34	14
134	Alaska-Ottawa	44	12	158	Utah-Florissant	18	10
134	Alaska-Harvard	48	15	1	Texas-Chicago	17	8
64	Alaska-Technology	49	14	1	Texas-Florissant	14	$12^1/_2$
11	Alaska-Charlottesville	48	12	1	Texas-Georgetown	24	12
134	Alaska-Florissant	42	13	3	Mexico-Florissant	23	12
11	Alaska-Florissant	42	13	3	Mexico-San Juan	36	13
11	Alaska-St. Louis	42	10	156	Nevada-San Juan	50	13
134	Alaska-Georgetown	48	14	1	Texas-Sitka	34	7
23	New Foundland-Victoria	45	14	156	Nevada-Sitka	22	$13^1/_2$
1	Texas-Charlottesville	23	11	3	Mexico-Sitka	44	11
156	Nevada-Ottawa	32	12	136	Alaska-Berkeley	30	10
1	Texas-Ottawa	27	10	138	Alaska-Berkeley	28	9
156	Nevada-Georgetown	32	12				
156	Utah-Georgetown	28	$11^1/_2$	11	Alaska-Honolulu	40	9
158	Utah-St. Louis	18	7	134	Alaska-Honolulu	41	9
156	Nevada-St. Louis	22	10	49	Chile-Wellington	83	$14^1/_2$
158	Utah-Ottawa	28	8	13	Hawaii-Victoria	38	$10^1/_2$
6	Mexico-Ottawa	38	10	169	Mexico-Honolulu	58	15

170	Mexico-Honolulu	50	14
171	Mexico-Honolulu	50	$12^1/_2$
40	Chile-Christchurch	94	14
40	Chile-Wellington	94	15
35	Central Atlantic-San Juan	56	13
46	East Atlantic-San Juan	47	$14^1/_2$
76	Off Africa-Georgetown	71	15
76	Off Africa-Harvard	70	12
35	Central Atlantic-Technology	67	16
94	S. Atlantic-La Plata	33	15
96	S. Atlantic-La-Plata	40	$14^1/_2$
96	S. Atlantic-Rio de Janeiro	42	13
95	S. Atlantic-Rio de Janeiro	43	11;14
94	S. Atlantic-Rio de Janeiro	39	15
42	Centr. Atlantic-Scoresby Sd.	60	$12^1/_2$
46	North Atlantic-Scoresby Sd.	32	11
32	North Atlantic-Scoresby Sd.	16	7
32	North Atlantic-Ivigtut	8	7
96	South Atlantic-Scoresby Sd.	124	15
106	SW. Indian Ocean-Wellington	86	$14^1/_2$
107	SW. Indian Ocean-Wellington	86	$14^1/_2$
107	SW. Indian Ocean-Melbourne	69	14
107	SW. Indian Ocean-Christchurch	83	14
107	SW. Indian Ocean-Batavia	53	16
107	SW. Indian Ocean-Colombo	46	15
106	SW. Indian Ocean-Bombay	53	13
107	SW. Indian Ocean-Bombay	54	$13^1/_2$
108	SW. Indian Ocean-Bombay	54	$13^1/_2$
134	Alaska-Scoresby Sd.	45	12
134	Alaska-Abisko	51	$15^1/_2$
138	Alaska-Abisko	53	$15^1/_2$
136	Alaska-Abisko	56	16
134	Alaska-Upsala	58	$15^1/_2$
24	Colombia-Rio de Janeiro	43	9
27	Peru-Rio de Janeiro	38	13
25	Ecuador-Rio de Janeiro	41	$10^1/_2$
49	Chile-Rio de Janeiro	28	10
25	Ecuador-La Plata	38	11
27	Peru-La Plata	34	12
37	Peru-La Plata	24	9
24	Colombia-La Paz	25	9
50	Chile-La Paz	19	10
96	S. Atlantic-Wellington	84	14
94	S. Atlantic-Wellington	78	14
94	S. Atlantic-Christchurch	76	15
94	S. Atlantic-Perth	83	15

7*

Table 16. Prevailing periods (seconds) in the cauda of the Sumatra earthquakes, 1931. (I = Febr. 10, II = Sept. 25)

Station	Dist. degr.	Periods I	Periods II
Batavia	4.4	9	9
Amboina	25.6	13 1/2	13 1/2
Colombo	25.6	13	13
Phu Lien	26.0	12 1/2	12
Manila	26.7	11 1/2	11 1/2
Hongkong	29.4	13	14 1/2
Perth	29.6	11	11
Calcutta	30.7	12	13
Bombay	37.7	11 1/2	13
Zikawei	40.2	13	14
Adelaide	44.6	?	14
Kobe	50.1	13; 17	14 1/2; 17
Melbourne	50.5	14	13
Riverview	53.4	11; 14	14
Tokyo	53.5	15	15 1/2
Vladivostok	54.8	13	?
Tananarive	55.3	15	15
Sverdlovsk	70.5	17	?
Wellington	73.3	17	16 1/2
Ksara	73.7	16	16
Johannesburg	73.9	18	?
Helwan	76.4	15	16
Capetown	82.3	15	?
Pulkovo	85.5	16	17
Tartu	86.9	17	?
Beograd	87.9	?	16
Helsingfors	88.2	17	17
Budapest	89.3	16 1/2	15
Wien	91.2	16	?
Zagreb	91.2	16	16 1/2
Graz	91.7	16	16 1/2
Upsala	91.8	?	17
Abisko	92.3	15	15
Potsdam	93.6	16	16
Lund	93.7	?	18
Eger	93.9	16	16
Leipzig	94.0	17	16 1/2
København	94.2	16	17
München	94.3	16	17
Firenze	94.3	17	17
Innsbruck	94.4	17	17
Jena	94.5	14; 17	15 1/2
Göttingen	94.5	16 1/2	16
Hamburg	95.6	17	16
Ravensburg	95.7	16	15
Chur	95.7	17	17
Stuttgart	96.0	17	17
Taunus	96.5	17	16
Karlsruhe	96.5	?	15
Strasbourg	96.9	?	16
Neuchâtel	97.5	16 1/2	17
De Bilt	98.5	17	16 1/2
Uccle	99.0	16 1/2	17
Paris	100.4	17	17 1/2
Alger	100.6	16 1/2	16
Honolulu	100.7	16	16; 19
Barcelona	101.1	?	15 1/2
Kew	101.9	17	17
Tortosa	102.4	?	16
Oxford	102.5	?	17
Edinburgh	102.9	?	17
Liverpool	103.3	19	15
Scoresby Sund	105.7	17	17
Cartuja	105.9	17	18
Toledo	105.9	16 1/2	17
Sitka	111.2	17	17
Ivigtut	119.8	?	16 1/2
Victoria	121.8	16 1/2	15-19
Saskatoon	127.2	18	18
Berkeley	127.8	?	18
Santa Clara	128.2	18	17
Bozeman	130.1	17 1/2	17 1/2
Pasadena	132.4	17 1/2	20
Rio de Janeiro	136.6	16	16
Seven Falls	137.7	17	?
Shavinigan Falls	138.4	?	16
Tucson	138.7	?	16
Halifax	138.7	?	18
Ottawa	139.7	17	18
Toronto	141.5	18	16
Buffalo	142.2	18	19
Chicago	142.3	19	17 1/2
Cambridge	142.4	?	16
Fordham	144.2	?	17
Georgetown	146.3	17	?
La Paz	156.7	17	?

Table 17. Prevailing periods (seconds) in the cauda of the Solomon Islands earthquakes, October 3, 1931. (I = Main shock, II = strongest after shock).

Station	Dist. degr.	Periods I	Periods II	Station	Dist. degr.	Periods I	Periods II
Riverview	25	11±	?	Helsingfors	121	$16^1/_2$	?
Apia	26	7±	?	La Plata	121	14	?
Guam	29	7	?	Shavinigan Falls	122	15	?
Adelaide	32	10	11	Fordham	123	?	16
Wellington	33	12	?	Seven Falls	123	16	?
Butuan	41	12	?	Capetown	124	17	?
Manila	48	11	?	La Paz	124	16	$15^1/_2$
Tokyo	51	12; 16	12	Upsala	124	16	?
Honolulu	51	9; 12	11; 14	Cambridge	124	$15^1/_2$	$15^1/_2$
Kobe	52	14; 17	?	Ivigtut	125	16	?
Hawaii Volc.	52	$11^1/_2$	?	Ksara	126	18	?
Sendai	52	14	?	Königsberg	126	15	?
Batavia	54	12	?	Halifax	129	17	?
Zikawei	57	15	14	Lund	129	$17^1/_2$	?
Hongkong	57	14	$14^1/_2$	København	129	$16^1/_2$	?
Phu Lien	63	16	?	Potsdam	131	17	?
Colombo	83	15	?	Hamburg	132	16	?
Sitka	85	15	$15^1/_2$	Budapest	132	17	16
Ukiah	85	$14^1/_2$	15	Leipzig	132	$16^1/_2$	16
Stanford	86	16	?	Beograd	132	$17^1/_2$	?
Berkeley	86	16	16	Wien	133	16	?
Lick	86	$15^1/_2$	?	Jena	133	17	?
Santa Barbara	87	$15^1/_2$	15	Göttingen	133	17	?
Pasadena	88	16	16	San Juan	133	?	$17^1/_2$
Victoria	88	15	15	Eger	133	17	?
La Jolla	88	16	?	Edinburgh	133	17	?
Riverside	89	$15^1/_2$	?	Zagreb	134	17	?
Haiwee	89	15	?	Taunus	135	18	?
Tinemaha	89	16	?	München	135	17	?
Bombay	92	15	?	Stonyhurst	135	16	16
Tucson	93	$15^1/_2$	16	Stuttgart	136	$17^1/_2$	?
Bozeman	96	16	15	Liverpool	136	17	?
Denver	99	$17^1/_2$	?	Innsbruck	136	17	?
Saskatoon	99	17	?	Karlsruhe	136	17	?
Tacubaya	102	16	?	Uccle	136	17	?
Tananarive	109	15	?	Ravensburg	136	18	?
St. Louis	111	16	?	Oxford	137	19	?
Chicago	113	15	15	Zürich	137	$17^1/_2$	?
Toronto	118	15	?	Paris	138	15	15
Columbia	119	$16^1/_2$	?	Rio de Janeiro	139	15; 18	?
Scoresby Sund	120	18	19	Barcelona	144	$17^1/_2$	?
Ottawa	120	16	$15^1/_2$	Toledo	148	$17^1/_2$	?
Georgetown	121	17	16	Cartuja	150	17	?

imum. This period may be much affected by conditions in the uppermost crustal layers, and may nevertheless be the same for two structures of completely different type.

Considering all the available data, the following general conclusions seem to be indicated. If the periods recorded at a given distance on different paths are compared, we find that the shortest periods are found in Eurasia and the North Atlantic, especially when the initial part of the path is through these regions. In the Central Atlantic the periods are nearer the average, while for shocks in the South Atlantic as recorded in South America the periods are decidedly above average. Wave paths crossing only North or South America result in periods near the average, as in Group IIa of Fig. 5. While in the northern or central part of the Pacific Ocean average or high periods prevail, there are quite numerous observations of paths through the southern Pacific Ocean with relatively short periods. The evidence as to the Indian Ocean is chiefly from shocks originating near 34° S. 57° E., which show decidedly long periods as recorded in India and at Batavia, but distinctly short periods at the stations in Australia and New Zealand. It must be remembered that in all cases the evidence is somewhat fragmentary, as the available paths cover only a small part of the areas studied. The problem calls for further investigation.

IV. Amplitudes of surface waves.

For all seismograms of the Sumatra and Solomon Islands earthquakes, where the necessary instrumental constants were available, the amplitudes of earth motion in microns were computed for the G waves, the maxima with periods from 16 to 22 seconds, and the corresponding waves of the W_2 and W_3 groups. The computations have been made from the usual formulas for continuous sinusoidal waves. The results are given in Tables 18 and 19. The stations are separated into regional groups, so that the effects of distance and azimuth are apparent on inspection. In the following subsections various types of conclusions from the data of these tables are discussed.

a) Effects of path and distance.

The amplitudes of surface waves are affected during their propagation by a number of factors, such as modification of the wave form, reflection and refraction at vertical discontinuities, absorption, etc. The first of these has not been studied thus far with the use of actual seismograms. A large loss of energy due to reflection and refraction at the boundary of the Pacific Basin has been suggested by GUTENBERG[23]. The data of Tables 18 and 19 provide two clear instances of this phenomenon. For the Solomon Islands shock (Table 18), the maximum waves arriving at Manila, Tokyo, Kobe, and Sendai have amplitudes between 1000 and 5000 microns; those arriving at Batavia, Zikawei, Hongkong, and Phu-Lien, at only slightly larger distances, have amplitudes from 250 to 700 microns. The difference is apparently due to loss of energy in passing the boundary between the regions of Pacific structure and continental structure. Apparently only a small fraction of the energy is transmitted across this boundary. That this is not due to instrumental conditions is shown by the facts that the G waves at Tokyo and Kobe have amplitudes of the same order as those at Hongkong, and that the maxima of the Sumatra shocks (one of which occurred only a few days before the Solomon Islands shocks) show almost the reverse relation of amplitudes at the two groups of stations.

For the Sumatra shocks (Table 19) the maximum amplitudes for the stations near the west coast of the United States (Ukiah, Berkeley, Santa Clara, Pasadena, Riverside, Tucson) are considerably smaller than those for most of the other North American stations (compare the immediately preceding and following groups of the Table). Reference to Fig. 3 will show that for the first-mentioned group of stations, which show the smaller amplitudes, the paths cross the boundary of the Pacific basin, while at the other stations this is not the case. Neither the boundaries of the Atlantic Ocean nor those of the Indian Ocean give evidence of a similar effect. These results are in good agreement with those found from the amplitudes of PP[2] and from the velocities of surface waves (section II of the present paper); but there is no indication of any effect associated with the Arctic basin.

The effect of absorption is expected to be of the form

$$(1)\qquad \frac{a_2}{a_1} = \frac{T_2}{T_1}\, e^{-k(\Delta_2-\Delta_1)/2} \sqrt{\frac{\sin\Delta_1}{\sin\Delta_2}}\ \sqrt[4]{\frac{\Delta_1}{\Delta_2}}$$

except for waves having periods near that for which the group velocity is a minimum, in which case the form

$$(2)\qquad \frac{a_2}{a_1} = \frac{T_2}{T_1}\, e^{-k(\Delta_2-\Delta_1)/2} \sqrt{\frac{\sin\Delta_1}{\sin\Delta_2}}\ \sqrt[6]{\frac{\Delta_1}{\Delta_2}}$$

Table 18.

Amplitudes in microns of surface waves in the Solomon Islands shock of October 3, 1931, 19^h. (I) and the strongest aftershock at 22^h (II). G are the long waves with periods over $^3/_4$ minute, M the maxima with periods between 16 and 22 seconds. G_2 and M_2 are the corresponding waves over the larger arc, G_3 are G-waves returning after a circuit of the earth.

Station	Dist. degr.	Shock I M	W_2	G	G_2	G_3	M/P	G/SS	G/M	II M	M_{II}/M_I
Riverview .	25.0	≧2000		5000					≦$2^1/_2$		
Apia	26.1	6500					22			1100	0.17
Adelaide . .	32.0	2800					90			600	0.21
Honolulu . .	50.9	2500		4000			31			350	0.14
Hawaii Volc. .	51.9	2000					80				
Manila . . .	47.5	1300					90			350	0.27
Tokyo . . .	50.6	4500		2000			65	3.3	0.4	700	0.17
Kobe	51.6	3500		5000			21		1.4	350	0.10
Sendai . . .	52.4	3500					120				
Batavia . .	54.4	250					5			70	0.28
Zikawei . . .	56.7	330					17			60	0.18
Hongkong . .	56.9	700		4000			14	20	6	80	0.11
Phu Lien . .	62.4	250								35	0.14
Calcutta . .	79.0	350					35				
Colombo . .	83.2	210		1000			21		5	60	0.29
Bombay . .	92.3	100		1000	600	50	17		10	30	0.30
Sitka	84.8	300		3000	600		60	20	10	100	0.33
Ukiah . . .	85.3	450		1100	700				$2^1/_2$	250	0.55
Stanford . .	85.6	200								60	0.30
Berkeley . .	85.7	450	30				23			60	0.13
Lick Obs. . .	86.0	250								45	0.18
Sa. Barbara .	86.7	450								120	0.27
Pasadena . .	87.9	450	25				45			100	0.22
Mt. Wilson .	88.0	120		500					4	20	0.17
Victoria . . .	88.2	250	15	3000	1900	300	17	5	12	40	0.16
La Jolla . . .	88.5	350		800					2.3	50	0.14
Haiwee . . .	88.5	300		1500					5	35	0.12
Tinemaha . .	88.6	650								120	0.18
Tucson . . .	93.5	150	30	1300	1100	150	30	4.6	8.7	70	0.47
Bozeman . .	95.6	500		1800			125	9	3.6	130	0.26
Denver . . .	99.3	1500		1200				$2^1/_2$	0.8	250	0.17
Saskatoon . .	99.4	600		2500					4.2	350	0.58
Chicago . . .	112.5	900		1300	900	80		$6^1/_2$	1.4	200	0.22
Toronto . .	118.1	300		1700	700			6.8	5.7	80	0.27
Pittsburgh .	118.4			2500				8.3		200	
Columbia . .	118.5	500		1000	1700	120		5	2	130	0.26

Table 18 (continued).

Station	Dist. degr.	Shock I M	W_2	G	G_2	G_3	M/P	G/SS	G/M	II M	$\frac{M_{II}}{M_I}$
Charlottesv. .	119.9	500		3000				15	6	300	0.60
Ottawa . . .	120.2	250		1500	300	60		$7^1/_2$	6	80	0.32
Georgetown .	120.6	200								30	0.15
Shavinigan F.	121.8			1700						130	
Seven Falls .	122.9	250		2000					8	50	0.20
Harvard . .	124.3	400		2000		300		6.8	5	100	0.25
Halifax . . .	128.5	400		500				3	1.3	120	0.30
La Plata . .	121.2	80		700	550			2.8	$8^3/_4$	10	0.13
La Paz . . .	123.8	100	30	600	900			1.2	6	15	0.15
Rio de Janeiro	138.7	200	40	1200	400			6	6	25	0.13
San Juan . .	133.2	200		600				1.2	3	80	0.40
Tananarive .	108.7	150		400				5	2.7	30	0.20
Capetown . .	123.5	100		1400					14	30	0.30
Ksara. . . .	125.5	300		700	400	70		1.8	2.3	25	0.08
Helwan . . .	130.1	70		200				5	3	30	0.43
Ivigtut . . .	125.0	130		1500				3.8	11	60	0.46
Sverdlovsk .	104.9			500	120						
Pulkovo. . .	119.1	700									
Helsingfors .	121.1	350								80	0.23
Upsala . . .	124.1	200		600				1.5	3	60	0.30
Königsberg .	126.3	400		500				3	1.3	120	0.30
Lund	128.7	600		900				1.3	$1^1/_2$	90	0.17
København .	129.4	450	10	500				1.2	1.1	100	0.22
Potsdam . .	131.2	600		200				0.8	0.3	60	0.10
Hamburg . .	131.6	500								150	0.30
Budapest . .	131.6	500								100	0.20
Leipzig . . .	132.3	250								60	0.24
Beograd. . .	132.4	1000								300	0.30
Wien	132.6	350		500					1.4	90	0.26
Jena	132.9	300								40	0.13
Göttingen . .	133.1	400		400				0.8	1.0	50	0.13
Eger	133.2	300	20		700					30	0.10
Edinburgh .	133.3	200		700				1.4	$3^1/_2$	70	0.35
Zagreb . . .	134.3	500		1000				1.8	2	80	0.16
De Bilt . . .	134.5	700		400				1.0	0.6	140	0.20
Taunus . . .	134.8	300									
München . .	135.0	550		400				0.5	0.7	60	0.11
Stonyhurst .	135.1	300		600				1.5	2	60	0.20
Heidelberg .	135.3	300		600				3	2	20	0.08
Stuttgart . .	135.6	250		550				1.1	2.2	50	0.20
Hohenheim .	135.6	250								70	0.28
Liverpool . .	135.6	350		400				0.5	1.2	70	0.20

Table 18 (continued).

Station	Dist. degr.	Shock I M	W_2	G	G_2	G_3	M/P	G/SS	G/M	II M	$\frac{M_{II}}{M_I}$
Innsbruck .	135.7	250								50	0.20
Karlsruhe . .	135.7	300									
Uccle . . .	135.8	400		600				$1^1/_2$	1.5	80	0.20
Ravensburg .	136.1	300		600				$1^1/_2$	2	40	0.13
Oxford . . .	136.7	200								60	0.30
Kew	136.7	450		300				$^3/_4$	0.7	75	0.17
Zürich . . .	137.0	350		300				1	0.8	50	0.14
Neuchâtel . .	137.9	350								60	0.17
Paris	138.1	350								70	0.20
Moncalieri . .	139.1	350									
Barcelona . .	144.4	250		600				1	2.4	25	0.10
Tortosa . . .	145.6	150		500				0.7	3.3	25	0.17
Alger	147.6	250		450				0.9	1.8	20	0.08
Toledo . . .	148.2	300		650				1.1	2.2	30	0.10
Cartuja . . .	150.4	250								30	0.12
San Fernando	152.0	350									

Table 19.

Amplitudes in microns of surface waves in the Sumatra shocks of Fe ruary 10, 1931 (I) and September 25, 1931 (II). G are the long wav with periods over $^3/_4$ minute, M the maxima with periods between and 22 seconds, W_2 the corresponding waves over the larger arc.

Station	Dist. degr.	Shock I G	Shock I M	Shock I W_2	Shock II G	Shock II M	Shock II W_2	II/I $\frac{P_{II}}{P_I}$	II/I $\frac{M_{II}}{M_I}$	II/I $\frac{W_{II}}{W_I}$	Shock I M/P	Shock I $\frac{M}{W_2}$	Shock M/P
Colombo . . .	25.6		600			900			1.5		60		
Calcutta . . .	30.7					1400							70
Bombay . . .	37.7		200			800		2	4		50		100
Phu Lien . .	26.0	1200	600			1000		2.1	1.7		50		40
Manila . . .	26.7		1600			1700			1.1				
Hongkong . .	29.4		500			1800		5	3.6		50		36
Zikawei . . .	40.2		350		400	500		1.7	1.3		29		25
Vladivostok .	54.8		100										
Kobe	50.1		150			250		2.5	1.7		19		13
Tokyo	53.5		350			500			1.3				
Sendai . . .	55.8		80			100		2.1	1.3		7		4
Perth	29.6		> 1400			> 1000		1					
Adelaide . . .	44.6					500							50
Melbourne . .	50.5		1000			> 2000		1	>2		100		
Riverview . .	53.4	500	700			700		1.5	1.0		350		350
Wellington . .	73.3		100			120			1.2				

Table 19 (continued).

Station	Dist. degr.	Shock I G	Shock I M	Shock I W_2	Shock II G	Shock II M	Shock II W_2	II/I $\frac{P_{II}}{P_I}$	II/I $\frac{M_{II}}{M_I}$	II/I $\frac{W_{II}}{W_I}$	Shock I M/P	Shock I $\frac{M}{W_2}$	Shock II M/P	Shock II $\frac{M}{W_2}$
Tananarive .	55.3		250			250			1.0					
Capetown . .	82.3		150											
Ksara	73.7		75			450		0.7	6		11		90	
Helwan . . .	76.4					> 100		5						
Irkutsk . . .	57.1		200			≧ 250			≧1.3					
Pulkovo . . .	85.5		100			≧ 300			≧3					
Kucino . . .	80.2		60			250			4.2					
Beograd . . .	87.9		20			130		2.3	6 1/2		7		19	
Helsingfors. .	88.2		70			200		2	2.9		23		33	
Budapest . .	89.3		35			170		4 1/2	4.7		18		18	
Wien	91.2		40								13			
Zagreb . . .	91.2		50			110		0.7	2.2		7		22	
Graz	91.7		50			120		1 1/2	2.4		25		40	
Upsala . . .	91.8					200					200			
Abisko. . . .	92.3		60			120		1 1/4	2.0		15		24	
Potsdam . .	93.6		70			220		1.7	3.1		23		44	
Lund	93.7					200							67	
Eger	93.9		70		150	150	10		2.1					15
Leipzig . . .	94.0		60			160			2.7				40	
København .	94.2	70	90	20		220	12	2	2.4	0.6	45	4 1/2	55	18
München . .	94.3		55			170		3	3.0		55		57	
Innsbruck . .	94.4		30			80		1	2.7		60		160	
Jena	94.5		20			70			3.5				70	
Göttingen . .	95.5		40		160	120			3				60	
Hamburg . .	95.5		200			350		1	1.8		67		117	
Ravensburg .	95.7		25			125		2	5		50		125	
Chur	95.7		50			150		4	3		100		75	
Hohenheim .	96.0		10			110			11				110	
Stuttgart . .	96.0		40	15		110	10	2	2.8	0.7	26	2.7	37	11
Heidelberg .	96.3		25			70			2.8				70	
Zürich . . .	96.4		25								50			
Karlsruhe . .	96.5					90							90	
Strasbourg . .	96.9					150								
Neuchâtel . .	97.5		45			150			3.1				300	
De Bilt . . .	98.5		150		100	300		1 1/2	2.0		150		200	
Uccle	99.0	70	70		80	200			2.9				100	
Paris	100.4		70			120			1.7				60	
Alger	100.6		40			90			2.3					
Barcelona . .	101.1		20			100			5.0					
Kew	101.9		90		100	230			2.6				115	
Tortosa . . .	102.4					70								
Oxford . . .	102.5		30			140			4.7					

Table 19 (continued)

Station	Dist. degr.	Shock I G	Shock I M	Shock I W_2	Shock II G	Shock II M	Shock II W_2	II/I $\frac{P_{II}}{P_I}$	II/I $\frac{W_{II}}{W_I}$	II/I $\frac{W_{II}}{W_I}$	Shock I M/P	Shock I $\frac{M}{W_2}$	Shock II M/P	Shock II $\frac{M}{W_2}$
Edinburgh .	103.0		45			110			2.4				55	
Liverpool . .	103.3		45											
Toledo . . .	105.9		35			65		1.3	1.9		70		100	
Scoresby Sd. .	105.7		40		90	100	10		2.5				100	10
Reykjavik . .	109.0					100								
Ivigtut . . .	119.8				160	200								
Honolulu . .	100.7		70	30		90	30		1.3	1.0		2.3		3
Sitka	111.2		40			100			2.5					
Victoria . . .	121.8		40	45		150	40		3.8	0.9		0.9		$3^3/_4$
Saskatoon . .	127.2		70			160			2.3					
Bozeman . .	130.1				230	230								
Ukiah . . .	126.5					50								
Berkeley . .	127.7		25			45	25		1.8					1.8
Santa Clara .	128.2		30	50		25	25		0.8	0.5		0.6		1.0
Pasadena . .	132.4	30	25	20		40	20		1.6	1.0		1.3		2
Riverside . .	133.1					35								
Tucson . . .	138.7		30	45	120	90	35	2*	3	0.8		0.8		2.6
Denver . . .	137.5					150								
Seven Falls .	137.7	170	90											
Shavinigan F.	138.4		50			150			3					
Halifax . . .	138.7		20			50			$2^1/_2$					
Ottawa . . .	139.7	80	90	60	200	170	30	1	1.9	0.5		1.5		5.7
Toronto . . .	141.5		55	55		140	90	1.3	2.5	1.6		1.0		1.6
Buffalo . . .	142.2							3	3					
Chicago . . .	142.3		80			250		1.3	3.1					
Harvard . .	142.4		75			220		1.1	2.9					
Georgetown .	146.3		100			120		2	1.2					
Charlottesv. .	147.1		150			250		2	1.7					
Columbia . .	151.0		55			100		5	1.9					
San Juan . .	162.7	70	170			90		1.3	0.5					
Pt. au Prince .	165.6					60								
Balboa H. . .	175.6								1					
Rio de Jan. .	136.6		90	25		70	40	1.2	0.8	1.6		3.6		1.8
La Paz . . .	156.7	110	110			50	50	1	0.5					1.0

* This and the following values refer to the ratio of the P'-waves.

should apply[21]). These formulae are based on investigations by JEFFREYS[24]). a_1 and a_2 are the amplitudes at distances Δ_1 and Δ_2 respectively; T_1 and T_2 are the corresponding periods; and k is the coefficient of absorption, which is taken as constant along the paths. On account of the last assumption, care should be taken to avoid using data for paths which cross the boundary of the Pacific basin; paths which lie wholly within or wholly without that area supply comparable data. Even in these cases it must be considered that k as determined from the observations is not simply a coefficient of absorption, but also contains the effect of change of wave form, and other factors, so that the coefficient of absorption in the strict sense is smaller than the empirical value of k. The formulae cannot be applied to stations close to the epicenter or to its antipodal point, as they presuppose that all the energy diverges from the epicenter and converges toward the antipodal point simultaneously, and with coherent phases, which would lead to infinite amplitudes at these points.

The value of k depends on the wave-length and the crustal structures traversed. The greater the wave-length, the less will be the influence of the upper crustal layers on the amplitudes, and consequently on k. Consonant with this conclusion is the fact, already mentioned, that the amplitudes of G are not much affected by the boundary of the Pacific Ocean; and the values of k for the G waves seem to be the same for all paths, within the limits of error. For the Sumatra shocks there are very few observations of amplitude for G. Using the amplitudes of G and G_2 at Eger, we find $k = 0.00006$. In the Solomon Islands shocks G and G_3 have been measured at eight stations; the values of k found from these data are between 0.00006 and 0.00012, with an average of 0.00008. These results are in fair agreement with the value 0.00012 found previously for G waves[25]).

It is more difficult to determine k with precision for the maxima, as k in this case is much more dependent on the upper crustal structures. For fairly accurate results rather long wave-paths must be used; the value of k found in this way is not a simple average or mean, since it may be influenced by high absorption over a short segment of the path. Thus the value of about 0.0003, which has been found for a complete circuit about the earth[25]), is much affected by passing the discontinuities; probably k is smaller than this in any one unit of the earth's crust. For waves with periods of about 18 sec. we find by using the amplitudes of M and W_2 in the large Solomon Islands shock, the following values of k: between South Africa (M) and North America (W_2), 0.00016; between Europe and South America (in either direction), 0.00032; between Apia and Hawaii (both M), 0.00028. While the first of these figures agrees well with previous results for the same regions [0.00017 for Eurasia, 0.00013 for the Atlantic Ocean, 0.00020 for America[25])], the second appears to be rather high. For the Pacific Ocean there is only one doubtful earlier result, of 0.00023; this agrees well with the result from the data of Apia and Hawaii, which itself is rather uncertain. For the Sumatra shocks we find the following results as averages between the two shocks (the individual data do not differ greatly): between Asia and North America across the Arctic region (both M), 0.00017; between Rio de Janeiro and North America (either direction) 0.00015; between Asia and Europe (M), 0.00011.

Table 20 gives amplitudes calculated from equation (1) for G, G_2, and G_3 assuming $k = 0.00008$, constant period, and unit amplitude for G at 10000 km.; also from equation (2) for M and W_2 assuming $k = 0.00016$, constant period, and unit amplitude for M at 10000 km. It should be noted that the amplitudes of the G group are not comparable with those of the M group, as no assumption has been made as the ratio of G to M; and that in the case of W_2 the true amplitudes will usually be smaller, due to loss of energy at discontinuities.

Table 20.
Calculated relative amplitudes of surface waves.

Distance	$k = 0.00008$			$k = 0.00016$	
km.	G	G_2	G_3	M	W_2
0	∞	∞	∞	∞	∞
3000	2.6	0.4	0.3	3.2	0.2
5000	1.7	0.3	0.2	2.0	0.1
7000	1.3	0.3	0.2	1.4	0.2
10000	1.0	0.3	0.1	1.0	0.2
13000	0.9	0.4	0.1	0.8	0.2
15000	0.9	0.5	0.1	0.7	0.3
17000	1.0	0.7	0.1	0.8	0.5
20000	∞	∞	∞	∞	∞

b) Amplitude ratios of several wave groups.

In the columns headed M/P in Tables 18 and 19 will be found the ratios of the amplitudes of ground motion in the maxima to those of the corresponding P groups. Considered as a function of distance, the ratio M/P shows a very wide scatter, which may partly be due to instrumental factors; but the results from the three shocks used agree reasonably well with each other, and show on the whole a behaviour similar to that of the ratio PP/P as given in the preceding paper[2]).

This is to be expected, as M varies with distance more slowly than P, and consequently the behaviour of M/P must largely be dependent on the amplitudes of P alone. The effect of M probably consists chiefly in a scattering of the values of the ratio, owing to the different amplitudes of M along different paths.

Between 20° and 90° the amplitudes of M are from 10 to 100 times those of the largest P wave. Assuming periods of 18 seconds in M and 6 seconds in P, the energy in a single wave in the maximum phase is a few hundred times that of a single P wave; assuming a period of 2 seconds in P, this ratio is divided by about 10. In the range where P is largest, near 60°, it may even happen that one P wave has the same energy as the largest M wave. Of course these results cannot be applied directly to determine the relative energies in the P and M groups, as the number of waves in each group must be taken into consideration.

In the large Solomon Islands shock both G and SS were conspicuous on account of their long periods, which are of the same order—roughly one minute—for both phases. At the North American stations, with distances ranging from 85° to 129°, G generally has several times the amplitude of SS; but at the European stations, distant 120° to 150°, the two phases are about equal. This is to be attributed to the behaviour of G, and not to SS (see below).

In this same shock, SS and M have amplitudes of the same order, so that the above remarks concerning G/SS apply equally well to G/M. The second Solomon Islands shock was characterized by comparatively small G waves; this phenomenon, and its bearing on the question of mechanism, have been discussed in our first paper[1]). In the two Sumatra shocks G and M are roughly equal.

c) Evidence of azimuthal effects due to mechanism.

For the principal Solomon Islands shock the calculated amplitudes of G in North America chiefly range from 1000 to 3000 microns. In Europe, on the other hand, the amplitudes are near 500 microns. The two ranges of distance overlap to some extent, and the large size of the discrepancy leaves no doubt that the result is actually due to a real distribution in azimuth. Since the G waves cannot be much affected by crustal structures along their path, this must be explained in terms of the mechanism of the shock.

For further investigation of such irregular distribution in energy we can investigate the ratio of the recorded amplitudes of the same wave groups in two shocks from the same source, or the azimuthal distribution of amplitudes of the same waves in a single shock.

For the two Solomon Islands shocks the amplitudes of the maxima show no clear distribution in azimuth; the ratios of the maximum amplitudes for the two shocks, as given in the last column of Table 18, are roughly constant with a mean value of 0.22. Corresponding mean ratios have been calculated for other phases; for P we find 0.25, and for S from a smaller number of measurements, 0.36. There are a few ratios available for P', PP, and SKP, which are all of the order of 0.2. The general constancy of these ratios indicates that the radiation of energy from the source was very similar in the two cases. Further, the fact that the amplitudes of M and W_2 are roughly constant in different azimuths at the same distance indicates that this radiation was rather uniform in both cases. However, as pointed out above, the G wave shows a definite distribution in azimuth, and is very much stronger in the first shock. These data are entirely consistent with the hypothesis put forward in section IV of our first paper[1]), that the circumstances of generation of these two shocks were largely the same in such respects as depth of focus, attitude of the fault-plane, direction of displacement, etc.; and that the difference between the two groups of seismograms must be attributed to a longer duration of the generating process in the first shock, which presumably implies a greater extent of the source in space, perhaps into greater depths as well as horizontally.

Conditions are quite different for the two Sumatra shocks. (Cf. Table 19.) While the ratio of amplitudes of the two P phases is fairly constant, with a mean of 2.0, the ratios of amplitudes of surface waves clearly depend on azimuth. These ratios lead to definite conclusions as to the radiation of energy in the two cases; in Table 19 the data are given in columns headed M_{II}/M_I and W_{II}/W_I, which are the ratios of M and W_2 for the two shocks, and M/W_2, which is the ratio of these two phases for each shock separately. There is an evident distribution in azimuth of the relatively large and small values for these ratios. In the northwest quadrant (including Europe), M_{II}/M_I is above average, while W_{II}/W_I is below average; in the same quadrant M/W_2 is small for shock I and large for shock II. These data are consistent, and clearly indicate that in this direction a larger proportion of energy was radiated in the maximum waves for the second shock than for the first. In the northeast quadrant M_{II}/M_I is large, W_{II}/W_I is small, but M/W_2 is small for both shocks; while in the southwest quadrant M_{II}/M_I is small, and M/W_2 is small for the second shock. (Data are rather scanty in the southwest quadrant, and there are no sufficient data in the southeast quadrant.) These results indicate that in both shocks, but more markedly in the first, a relatively small amount of energy was radiated in the maximum waves toward the northeast, while toward the southwest more than the average energy was radiated in the first shock, but less in the second. In this way it appears that, in spite of the fact that the hypocenters of these two shocks were nearly identical, the mechanism of origination differed in some way which produced a considerable difference in the azimuthal distribution of energy in the surface waves, although the P waves were not similarly affected.

V. On the magnitude and energy of earthquakes.

a) The magnitude of the Solomon Islands and Sumatra shocks, 1931.

In a recent paper [26]) one of us has described the construction and application of a magnitude scale for earthquakes in the California region. The term magnitude is here employed as referring to a quantity characteristic of the shock as a whole, in distinction from the intensity, which refers to its manifestation at a particular point. The magnitude is defined as the (Briggs) logarithm of the maximum registered trace amplitude measured on the seismogram, when expressed in units of 0.001 mm. (= 1 micron), as recorded by a standard short-period torsion seismometer ($T_0 = 0.8$, $V = 2800$, $h = 0.8$) at a distance of 100 km.

The reduction of observed amplitudes to expected amplitudes at the standard distance of 100 km. is accomplished by means of a tabulation of amplitudes for a standard shock at various distances. This tabulation was constructed from the recorded amplitudes of well-located shocks at the stations in Southern California. Such a procedure involves the assumption that the ratio of amplitudes at two given distances is the same for all shocks and in all azimuths. The method consequently yields only very rough results; it is of practical value only because the actual range in shock magnitudes is so enormous that even the crudest means are sufficient to separate them into a convenient number of levels. Fortunately, in the Southern California region the depths are fairly constant—mostly near 15 km.—so that uncertainties from this source are not important. For further discussion see the end of this section.

In the original paper the magnitude scale was extended only to distances of about 600 km., with some indication as to possible extrapolation beyond. Data have now become available which leave no considerable gaps up to about 1500 km. (about 13°), and make it possible to study the maximum registered amplitude as a function of distance. It is clear that this cannot be a single function over the whole range, since at the shorter distances the maximum of the seismogram is one of the S group of waves, while at the larger distances it is a surface wave. However, from 200 to 1500 km. the observed amplitudes are fairly well represented by the formula

$$\log A = \log M + 3.37 - 3 \log \Delta \tag{3}$$

where A is the seismographic amplitude in millimeters, M is the magnitude as defined above, and Δ is the distance in kilometers. At larger distances the magnitudes calculated from (3) appear too high, indicating that the amplitudes at these distances do not decrease as rapidly as the formula requires.

The data of the present paper offer a possibility of extending the magnitude scale to all distances. The observations are rather well represented by equation (2) for all available distance (25° to 150°), seldom being less than 0.3 or more than 0.3 times the calculated values, which corresponds to an error of less than half a unit on the magnitude scale.

The amplitude A, in millimeters, recorded by the standard torsion seismometer (constants as mentioned previously) is given by

$$A = \frac{V}{1000 \sqrt{(u^2-1)^2 + 4\, h_2\, u^2}}\, a \tag{4}$$

where V is the static magnification, $u = T/T_0$ (T = period of earth motion, T_0 = period of the pendulum), h is the damping constant, and a is the ground amplitude in microns. At the distances named the period of the maximum surface waves is large compared to the free period of the standard torsion seismometer; u is large compared to 1, and u^4 is large compared to $4\, h^2\, u^2$, so that (4) can be replaced by

$$A = \frac{V\, T_0^2}{1000\, T^2}\, a. \tag{5}$$

With the standard values of V and T_0 this gives

$$A = \frac{1.8}{T^2}\, a. \tag{6}$$

Introducing this result into (2), we find

$$\frac{A_2}{A_1} = \frac{T_1}{T_2}\, e^{-k(\Delta_2 - \Delta_1)/2} \sqrt{\frac{\sin \Delta_1}{\sin \Delta_2}}\, \sqrt[6]{\frac{\Delta_1}{\Delta_2}}. \tag{7}$$

Since k is small, the exponential may be neglected when $\Delta_2 - \Delta_1$ is not very large; and for distances under 40° $\sin \Delta_1 / \sin \Delta_2$ may be replaced by Δ_1/Δ_2, so that finally

(8) $$\frac{A_2}{A_1} = \frac{T_1}{T_2}\left(\frac{\Delta_1}{\Delta_2}\right)^{2/3}$$

which will apply with considerable accuracy for waves with periods over 4 seconds at short distances.

For short distances (200—1500 km.) the empirical formula[26] is $A_2/A_1 = (\Delta_1/\Delta_2)^3$. This is mainly due to the variation of period with distance, as observations at Pasadena indicate[16] that the periods in the surface waves increase from about $^1/_2$ second at 200 km. to about one second at 500 km. and 6 seconds at 800 km. For the larger of these distances this gives roughly an increase of the period with the square of the distance; and it is possible that this effect combines with the variation with the $^2/_3$ power of Δ in the previous formula (8) and the effect of absorption to give a variation of recorded amplitude approximately as the inverse cube of the distance. At greater distances this power should decrease.

Until the magnitude scale can be extended directly up to 25° or 30°, its connection with amplitudes at large distances involves a considerable extrapolation. This has been undertaken in the following way: (Method A) the amplitudes, at the shortest distances for which reliable observations are available, are reduced to amplitudes calculated for 15° by means of formula (2); the corresponding recorded amplitudes for the standard torsion seismometer are calculated from (6), assuming $T = 10$ sec.; the magnitude scale is extended to 15° by the inverse cube rule which leads to an expected amplitude of registration of 5.1×10^{-7} mm. for the standard shock (magnitude 0, see below), the logarithm of which is -6.29; the algebraic subtraction of the last quantity from the logarithm of the calculated amplitude for the shock being investigated gives its magnitude. A similar procedure can be applied to any other distance; at 10°, for example, the period of the maximum as taken as 7 seconds, and the logarithm of the expected registered amplitude for the standard shock is -5.77. At 15° the extrapolation from larger distances is probably better than that from shorter distance; while at 10° the reverse conditions apply.

The problem may be approached in a quite different way (Method B): For the larger shocks in the region of Pasadena, to which the magnitude scale has been applied, amplitudes of the maxima have been given in the bulletins of a number of stations. These data are given in Table 21. The horizontal amplitudes as given in microns are estimated or computed from the amplitudes of the several components as reported, due allowance being made for those cases in which only one or two components are available. These stations are at distances between 70°

Table 21.

Ground amplitudes (horizontal) in the maxima of surface waves with periods of about 18 seconds in microns.

	Dec. 21 1932 6^h	Mar. 11 1933 1^h	June 6 1932 8^h	June 25 1933 20^h	Jan. 30 1934 15^h	Mar. 12 1934 15^h	Dec. 30 1934 14^h	Dec. 31 1934 18^h
Hamburg .	350	20	30	7	40	60	50	250
Kew. . . .		30				40	50	200
Jena. . . .	140	5				15	20	200
Göttingen .	250	25	10	7	25	20	35	200
Stuttgart. .		25					40	150
Toledo . . .	350	20	8	7	5			
Cartuja . .	170	20	20	7	25			
Pulkovo . .	120	20	40	6		25	25	120
Kucino . .	110	10	15	6		20	45	100
Sverdlovsk .	50	15	15	5			25	60
Taschkent .		15	15	3		20		
Chiufeng . .	70	10				45	15	35
Vladivostok					20	40		25
Melbourne .		5						45
La Paz . .	350	10	15		15	30	30	400
Average . .	200	15	20	6	25	30	35	150
Magnitude .	7.5	6.2	6.1	6.2	6.5	7.0	6.1	6.5
log average.	2.3	1.2	1.3	0.8	1.4	1.5	1.5	2.2
Difference .	5.2	5.0	4.8	5.4	5.1	5.5	4.6	4.3

and 100°, within which range the effect of distance is not large, and can be allowed for by applying the results of Table 20. In the last lines of the table are given: the average ground amplitudes estimated for a distance of 90°; the magnitude assigned to the shock from the data of the Southern California stations; the logarithm of the assumed average amplitude, and the difference obtained by subtracting this from the magnitude. The last quantity should theoretically be a constant for all shocks, variations being due to irregularities in the radiation of energy, effects of wave paths, and numerous minor sources of error. These effects are especially evident in the first and last shocks tabulated.

From the last line, the mean value of the constant is about 5.0, resulting in the formula

$$M = 5.0 + \log a \text{ (for } \Delta = 90°). \tag{9}$$

This makes it possible to determine the magnitude M of any shock for which observations of the ground amplitude a are available. This extension of the scale neglects the effect of variations in depth (see the end of this section).

In the region of Pasadena it has been found that the smallest shocks recorded are of magnitude 0 (standard shock of the scale); the smallest shocks reported felt are of magnitude 1.5, while ordinarily perceptibility does not occur below 2.5; damage in the epicentral region begins to occur about magnitude 4.5; shocks destructive over a limited area are of magnitude 6; major earthquakes (recorded over the world, geological effects at the surface) exceed magnitude 7.

For the large Sumatra and Solomon Islands shocks we are fortunate in having available ground amplitudes determined from the original seismograms for a great number of stations. Accordingly, it becomes very interesting to apply methods A and B to find the magnitudes of these shocks.

For method A we can use for the Solomon Islands shocks only the data of Apia (distant 26°); for the Sumatra shocks we have used the mean of the data of Colombo, Phu-Lien, and Manila (all near 26°). The amplitudes at Apia are about 6500 and 1100 microns; the mean values used for the Sumatra shocks are 800 microns for the February shock and 1100 microns for the September shock. In all cases the per ods of these maxima are about 17 seconds. These data have been reduced to 10° and 15°, the magnitude scale being then applied as described above.

For method B we have taken the following mean amplitudes for a distance of 90°: Solomon Islands shocks, 400 and 100 microns; Sumatra shocks, 75 and 150 microns. The magnitudes are then found from formula (9).

The results are as follows:

	Method A (10°)	Method A (15°)	Method B
Main Solomon Islands shock	8.1	8.3	7.6
Aftershock	7.4	7.6	7.0
Sumatra, February . . .	7.2	7.5	6.9
Sumatra, September . . .	7.4	7.6	7.2

Method B should give better results, as it involves no extrapolation, and in this case uses the data of a much larger number of stations. Hence it appears that Method A applies better at 10° than at 15°; and this is to be expected, as the extrapolation of the magnitude scale to 15° by the inverse cube rule is uncertain; this rule is empirical, and cannot apply at the larger distances where the maxima certainly decrease more slowly with distance.

b) Preliminary extension of the magnitude scale to large distances.

The fact that the methods of the preceding paragraphs lead to consistent and reasonable results, indicates that it is possible to extend the magnitude scale to large distances. The following general procedure can be employed.

The magnitudes M of a certain number of strong shocks are determined from the average ground amplitudes in microns at distances of the order of 90°, by applying formula (9). Now if the amplitude A recorded on the standard torsion seismometer at distance Δ is known, we can determine the theoretically expected amplitude of registration of the standard shock (magnitude 0) at distance Δ, by using the relation

$$\log A_0 = \log A - M. \tag{10}$$

Table 22 lists the shocks used, giving the distances from Pasadena, the magnitudes determined as mentioned above, the measured amplitudes of registration A, and the calculated values of $\log A_0$. The logarithms are given as negative quantities (negative exponents of 10) and not in the conventional form with positive decimal parts. In Fig. 6 the values of $\log A_0$ are plotted as ordinates against Δ as abscissa; for Δ a logarithmic scale has been used in order not to crowd the data for short distances. The two points at 8° refer to the S and M phases (S higher) for the Eureka shock of June 6, 1932. The point at 2.6° is from the Parkfield earthquake of June 7, 1934. The full line at the shorter distances represents the data for the standard shock given in Table 1 of the previous paper on the magnitude scale[26]); the dashed line continuing up to 15° gives the extrapolation of these data by means of the inverse cube rule. The dashed line for distances over 20° is calculated from equation (2) above, assuming that the period does not change with distance, and that the coefficient of absorption k has the value 0.00036, which is the mean value for surface

Table 22.

Magnitude M of earthquakes (calculated from published amplitudes of their maxima); trace amplitudes A recorded by the short period torsion seismographs at Pasadena; $\log A_0 = \log A - M$.

No.	Date	Epicenter	Dist. degr.	M	A mm.	$\log A_0$
111	1932, July 12	Gulf of California	$11^1/_2$	6.7	$8^1/_2$	—5.8
1	1931, Aug. 16	Texas	$12^1/_4$	6.3	9	—5.4
142	1934, Sept. 15	Mexico	$17^3/_4$	6.2	1	—6.2
112	1931, Jan. 2	Mexico	$18^1/_2$	6.4	$4^1/_2$	—5.7
154	1934, Nov. 30	Mexico	$18^1/_2$	6.8	5	—6.1
113	1933, April 9	Mexico	$19^3/_4$	6.1	1	—6.1
—	1932, June 3	Mexico	20	7.9	90	—6.0
124	1933, Dec. 13	Mexico	20	6.5	2	—6.2
3	1932, Dec. 7	Mexico	20	6.6	3	—6.1
6	1933, July 10	Mexico	22	6.1	0.7	—6.3
126	1934, Jan. 28	Mexico	24	6.7	2	—6.4
7	1928, June 17	Mexico	$25^1/_2$	7.8	20	—6.5
8	1928, Oct. 9	Mexico	$25^1/_2$	7.5	6	—6.7
138	1934, May 14	Alaska	33	5.9	0.15	—6.7
134	1934, May 4	Alaska	34	6.8	1	—6.8
10	1932, May 21	Central America	34	7.0	$1^1/_2$	—6.8
11	1933, April 27	Alaska	34	6.9	$1^1/_2$	—6.7
136	1934, July 28	Alaska	$34^1/_2$	6.6	0.5	—6.9
140	1934, July 18	Panama	42	7.4	2	—7.1
22	1933, Nov. 20	Baffin Bay	46	7.5	16	—6.3
26	1929*), Dec. 17	Aleutian Is.	$51^1/_4$	7.7	4	—7.1
44	1933, March 2	Japan	75	8.3	7	—7.5
49	1931, March 18	Chile	80	7.0	0.5	—7.3
147	1934, July 18	Sa. Cruz Is.	$84^1/_2$	7.9	$1^1/_2$	—7.7
148	1934, July 21	Sa. Cruz Is.	$84^1/_2$	7.2	0.8	—7.3
57	1931, Nov. 2	S. Japan	87	7.4	$1^1/_2$	—7.2
59	1931, Oct. 10	Solomon Is.	88	7.5	$1^1/_2$	—7.3
—	1931, Oct. 3	Solomon Is.	88	7.6	0.6	—7.8
—	1929, June 16	New Zealand	94	7.5	1	—7.5
119	1934, March 5	New Zealand	96	7.1	0.5	—7.4
71	1931, Aug. 10	Altai Mts.	96	8.1	2	—7.8
73	1932, Dec. 25	Kansu	100	7.9	1	—7.9
85	1932, May 14	Celebes	111	7.8	0.5	—8.1
89	1934, Jan. 15	India	115	8.2	0.8	—8.3
97	1933, June 24	Sumatra	131	7.5	0.2	—8.2
—	1931, Sept. 25	Sumatra	$132^1/_2$	7.2	0.1	—8.2
107	1933, Jan. 21	Indian Ocean	176	7.0	0.7	—7.2

*) In Table 4 of the first paper[1]) misprinted as "1931".

waves which have crossed the boundary of the Pacific basin[25]). A dotted line has been drawn to represent the present observations at distances under 25°; the divergence between this and the calculated

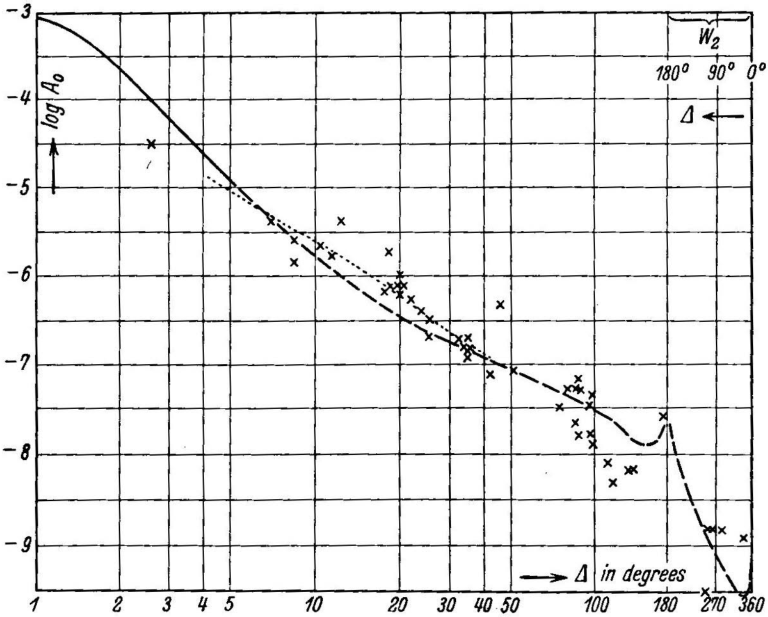

Fig. 6. Amplitude A_0 of a shock with the magnitude 0 as a function of the distance Δ. Abscissa: Δ in degrees, logarithmic scale. Ordinate: log A_0, where A_0 is the average amplitude of two horizontal components in mm. as recorded on the standard torsion seismometer.

Table 23. Log. A_0, calculated from W_2-waves.

No.	Date	Epicenter	Dist. degr.	M	A mm.	log A_0
89	1934, Jan. 15	India	245	8.2	0.05	—9.5
71	1931, Aug. 10	Altai Mts.	264	8.1	0.2	—8.8
—	1929, June 16	New Zealand	266	7.5	0.05	—8.8
44	1933, March 2	Japan	285	8.3	0.3	—8.8
—	1932, June 3	Mexico	340	7.9	0.1	—8.9

curve is principally due to the change in period with distance below 25°; this has the consequence that the magnification of the standard torsion seismometer for the maximum waves varies in this range, so that the registered trace amplitude decreases with distance more rapidly than the ground amplitude.

From 30° to 100° the agreement between the observations and the calculated curve is good. The points beyond 100° are about 0.3 unit lower than the curve, which means that the observed amplitudes are about half those excepted. This may be mere chance, as only a few shocks are available, and the registered amplitudes are very small. The very low point at 115° represents the observation for the Indian earthquake of Jan. 15, 1934; in this case the path is continental, so that amplitudes larger rather than smaller than the mean should have been expected. The point highest above the curve is that for the Baffin Bay shock of 1933 (46°).

The point at 176° (shock of Jan. 21, 1933) shows the relatively large amplitude near 180° which the theory requires. The points between 180° and 360° are taken from observations of W_2 at Pasadena (Table 23).

For the ground amplitudes at 90°, in microns, we have found the relation (9), which permits of calculating the shock magnitude when these are given. From the data of Fig. 6 we now find

$$M = 7.5 + \log A \text{ (for } \Delta = 90°) \tag{11}$$

where A is the measured amplitude in millimeters on the seismogram of the standard torsion instrument. The difference 2.5 between the constant terms in (9) and (11) should be the logarithm of the quantity $1.4\,T^2/1.8$, where T is the period of the maximum waves at 90°. This is taken from equation (6) for the magnification of the torsion instrument; the factor 1.4 arises from the circumstance that the components have been combined in estimating ground amplitudes, while the measured trace amplitude has been taken as the mean between the two horizontal components. For a period of 18 seconds the logarithm in question is 2.4; for 20 seconds it is 2.5, so that the results obtained in different ways, and based on different data, are in good agreement.

As has been mentioned, the material on which Fig. 6 is based, is not homogeneous; for short distances the maximum amplitude of the seismogram is one of the S group of waves, while for larger distances it is a surface wave. Moreover, there is a physical difference between shocks. In the California region it is already established that the large shocks usually show a higher energy concentration in the long-period maximum waves than to the smaller shocks, so that assigned magnitudes for the larger shocks have been based partly on the ratios of amplitudes in the preliminary part of the seismograms. In a number of cases, two different magnitudes can be assigned, according as the ratio of the amplitude of the large shock to that of a smaller shock is determined from the preliminaries or the maximum. In the neighborhood of 8° the seismograms show two definite maxima, one in the S phase, and the other considerably later in the surface waves; in this range of distances certain shocks show a predominance of short periods, others long periods [cf. our previous paper[2]), sect. IX—X]. In consequence of these various circumstances, there appear to exist two fairly well-defined curves near 8°; these may arise from seismograms of different character, or from different treatment of the same seismograms.

A further check on the methods used is provided by the two Mexican shocks of 1932 July 7, 16^h, and July 12, 19^h, respectively distant 7° and $11^1/_2$° from Pasadena. For these the magnitudes have been determined from the data of the Southern California stations (extrapolating the previous magnitude scale by the inverse cube rule) to be 6.6 and 6.8; while the magnitudes determined from the reported ground amplitudes at eleven distant stations, using formula (9), are 6.7 and 6.6. The corresponding values of $\log A_0$ have also been plotted in Fig. 6.

The observations of Fig. 6 can be represented reasonably well by the empirical relation

$$(12) \qquad \log A_0 = -3.7 - 2 \log \Delta, \text{ or } A_0 = 1/5000\, \Delta^2$$

where A_0 is given in millimeters and Δ in degrees.

The previous results make it possible to determine an approximate magnitude for a given shock from the data of a single station. Where torsion seismometers are available, we apply the general relation

$$(13) \qquad M = \log A - \log A_0$$

where A_0 is taken from Fig. 6. In other cases, for distances beyond 25°, when the horizontal ground amplitude a is known, we have

$$(14) \qquad M = \log a - \log A_0 - 2.5.$$

In these formulas $\log A_0$ is a negative quantity.

When the data of a number of stations are available, especially if well distributed in azimuth, the magnitude can be determined reasonably well.

The bulletins of a considerable number of seismological stations for the past thirty years have been examined with a view to identifying and determining approximate magnitudes for the very largest shocks. The results are exhibited in Table 24. It is believed that the list is nearly complete for shocks of magnitude 8 and over; however, the material is not homogeneous. For certain years, such as 1904, 1905, 1915, only two or three station bulletins have been available. A number

Table 24.
Magnitudes of selected large shocks.

Date	Reg. of epicenter	*M*	Date	Reg. of epicenter	*M*
1904, June 25	Kamchatka	8	1919, Jan. 1	W. Tonga Is.	8
1905, April 4	Kangra, India	$7^3/_4$	1919, April 30	Tonga Is.	8
1905, July 9	SW of Lake Baikal	8	1919, May 6	Bismarck Is.	8
			1920, June 5	Formosa	$7^3/_4$
1905, July 23	SW of Lake Baikal	8	1920, Dec. 16	Kansu	8.5
			1921, Sept. 11	East Indies	$7^1/_2$
1906, Jan. 31	Colombia-Ecuador	$8^1/_2$	1922, Sept. 1	Formosa	8
			1922, Nov. 11	Chile (Atacama)	8.4
1906, April 18	San Francisco	$8^1/_4$	1923, Febr. 3	SE. of Kamchatka	8.3
1906, Aug. 17	Aleutian Is.	8			
1906, Aug. 17	Chile (Valparaiso)	$8^1/_4$	1923, Sept. 1	Japan	8.1
			1924, April 14	Philippine Is.	8.0
1906, Sept. 14	New Guinea	8	1924, June 26	SW of Macquarie Is.	7.7
1907, Jan. 4	Sumatra	8			
1907, April 15	Mexico (Chilpancingo)	$7^3/_4$	1927, March 7	Japan	7.6
			1927, May 22	Kansu	7.8
1907, Oct. 21	Karatag	$7^1/_2$	1928, June 17	Mexico	7.8
1910, Nov. 9	New Hebrides	$7^3/_4$	1928, Dec. 1	Chile	8.0
1910, Dec. 13	East Africa	$7^1/_2$	1929, June 27	South Atlantic	7.8
1911, Jan. 3	Turkestan	$8^1/_2$	1929, Dec. 17	Aleutian Is.	7.8
1911, Febr. 18	Ferghana	$7^3/_4$	1931, Jan. 15	Mexico	8.0
1911, June 15	China Sea	8	1931, Febr. 2	New Zealand	7.7
1911, Aug. 16	Caroline Is.	8	1931, Aug. 10	Altai Mts.	8.1
1912, May 23	Burma	8	1932, May 14	Celebes	7.8
1912, Aug. 9	Turkey	8	1932, June 3	Mexico	7.9
1915, Oct. 3	Nevada	$7^3/_4$	1933, March 2	Japan	8.3
1917, May 1	S. Tonga	8	1934, Jan. 15	India	8.2
1917, June 26	Tonga	$8^1/_4$	1934, July 18	Santa Cruz Is.	7.9
1918, Aug. 15	Caroline Is.	8	1935, May 30	Baluchistan (Quetta)	7.7
1918, Sept. 7	Kurile Is.	$7^3/_4$			

of shocks with calculated magnitudes slightly below 8 have been included, on account of the possibility that the true magnitudes are higher.

A list of this character differs considerably from most ordinary lists of strong shocks; one the one hand, it includes great shocks occurring in unpopulated regions or ocean basins, and on the other hand it omits shocks of lesser magnitude whose epicenters were so situated as to make them conspicuous by extensive damage or loss of life. An instance of the latter sort is the Messina earthquake of 1908, which has a calculated magnitude of about 7. It is noteworthy that the destructive New Zealand shock of 1931 is assigned 7.7; that of 1929 is assigned 7.5.

The largest shocks appear to be of magnitude about $8^1/_2$. There are four of these in the table. The first is that of January 31, 1906, in the coastal region of Ecuador and Colombia; this shock was destructive over a very wide area. Next is the Turkestan earthquake of January 3, 1911, in the northern Tien-shan near Issykul. The third is the Chinese shock of December 16, 1920, well-known for great destruction and loss of life. The fourth, the Chilean shock of November 11, 1922, has a calculated magnitude of 8.4, but the accuracy of determination does not warrant ranking it below those just mentioned. Even some of the shocks which have been assigned magnitude $8^1/_4$ may belong to this group. One of these is the San Francisco shock of 1906. This result is of considerable interest; the large magnitude is presumably connected with the great linear extent of faulting.

While no great precision can be claimed for the magnitudes given in Table 24, they are sufficient to show the inadequacy of any listing of large shocks on the basis of reported macroseismic effects. Any cataloguing which is to be of real use for statistical purposes must be based on instrumentally determined magnitudes rather than on intensities — particularly if such statistics are to be used for conclusions as to the forces producing or occasioning earthquakes.

In the first paper on the magnitude scale[26]) an attempt was made to correlate the numbers of the magnitude scale with values for the energies released in the various shocks. This was done on the assumption that the Nevada shock of 1932 (magnitude 7.5) was of the same magnitude as the Montana shock of 1925, for which an energy of 10^{21} ergs has been calculated by Jeffreys[27]). This would imply the very small energy of 10^6 ergs for the smallest shocks recorded (magnitude 0). The data now in hand allow us to revise this result. Using the data of distant stations for the Montana shock, and applying the methods outlined above, we find a magnitude of 6.8. This gives for a shock of magnitude 0 an energy E_0 between 10^7 and 10^8 ergs. Since the magnitude scale is logarithmic in the amplitudes, doubling the magnitude gives a scale logarithmic in the energies, so that

$$\log E_0 = \log E - 2\,M. \tag{15}$$

A similar use may be made of Jeffreys' result of 10^{21} ergs or slightly more for the Pamir earthquake of Feb. 18, 1911 [28]). Using the station reports we find a magnitude of about 7, which would give 10^7 ergs, or slightly more, for E_0.

Reid[29]) calculated an energy of 1.75×10^{24} ergs for the San Francisco earthquake. The magnitude of $8^1/_4$ for this shock, given in Table 24, results in a value of E_0 slightly less than 10^8 ergs.

Energies have been found for earthquakes by many investigators using different methods; the results for the largest shocks are nearly always between 10^{25} and $^1/_2 \times 10^{26}$ ergs. As the magnitude of the largest shocks appears to be about $8^1/_2$, we find E_0 between 10^8 and $^1/_2 \times 10^9$ ergs.

The fact that all these various data result in a value of E_0 near 10^8 ergs, supports the reality of the magnitude scale and the validity of the methods used in determining magnitudes. It appears that in the not distant future it may be possible to replace the arbitrary numbers of the magnitude scale by others more directly related to the energy of the shock; but the correlation is not sufficiently precise at present, and it does not seem advisable to take this step prematurely.

The value $E_0 = 10^8$ ergs (1 kilogram-meter) may appear rather small when considered as the energy of an actual recorded earthquake. However, shocks of this magnitude are the very smallest observed, being rarely registered on the most sensitive instruments at distances of the order of 15 km. In such cases the seismogram consists chiefly of a single wave (S?), so that the whole recorded disturbance is concentrated in a very small volume. Calculation using the constants of the instruments indicates that in such a case an original energy of the order indicated may give a record of this kind. With this value of E_0, the total energy of any shock of given magnitude may be estimated by using equation (15).

In applying the formulas and other relations developed in the preceding section, due attention must be paid to the uncertain nature of the assumptions. The relation between shock magnitude or energy and the observed amplitudes is a very complex one, being much affected by local structure, mechanism of the shock, depth of focus, instrumental factors, and the like. Consequently, no great precision is claimed for these methods, but it is believed that they offer a possibility of placing the whole problem of seismic energy on a more accurate and consistent basis, besides considerably reducing the labor necessary to arrive at useful conclusions.

In extending the magnitude scale to shocks in other regions than

California there is an implicit assumption that the conditions of shock occurrence, particularly the depth of focus, do not differ widely. There is reason to suppose that the normal depth of local shocks in California (about 15 km.) is less than that usual in some other regions. Accordingly, these shocks may be expected to give rise to larger surface waves than shocks of equal energy in other regions, which would lead to unduly high magnitudes for the California shocks, or unduly low magnitudes for others.

However, it is probable that in most large shocks the generating disturbance extends to or near the surface; and the actual effect of variation in depth may not be very large, as will appear from the following discussion.

In a homogeneous body the theory requires that the Rayleigh waves decrease exponentially with depth. The amplitudes are proportional to

$$e^{-0.78\,\pi\,h/\lambda}$$

where h is the depth of focus and λ the wave-length[35] [36]). A similar formula applies to Love waves in a body consisting of two homogeneous layers[36]). The numerical coefficient in this case depends on the conditions, and is usually slightly larger than that for Rayleigh waves.

On the other hand, as JEFFREYS[36]) has pointed out, the observations indicate that surface waves decrease more slowly with focal depth than the theory requires. This is especially apparent in the case of local shocks with short-period surface waves.

Assuming a wave-length of 40 km., a shock with focal depth 10 km. should give a magnitude 0.4 larger than an equal shock with focal depth 25 km., while a shock at 50 km. depth should give a magnitude 0.7 smaller than at 25 km. If the surface waves actually decrease more slowly with focal depth than the theory indicates, these errors are smaller. A possible cause for such a difference between theory and observation is the fact that the surface of the earth is not homogeneous; moreover, we may observe surface waves travelling in a discontinuity at a depth nearly equal to the depth of focus. This last possibility, which could occur only for the shorter waves, has been referred to in section II.

VI. Note on seismic sea waves.

The large Solomon Islands earthquake gave rise to a considerable sea wave, which was destructive on the neighboring island coasts, and was registered on tide gauges at considerable distances. A tracing of the mareogram at Santa Barbara, California, was kindly made available to us by the U. S. Coast and Geodetic Survey; that at Hilo, Hawaii, is reprocuded in The Volcano Letter, No. 361, Nov. 26, 1931, published by the Hawaiian Volcano Observatory. The times of arrival of the first waves measured on these records are 1931, Oct. 4, 3.5^h at Hilo, and 9.0^h at Santa Barbara. The travel times are consequently 8.3^h and 13.8^h respectively. The great-circle distances from the epicenter to these two stations are 5800 and 9630 kilometers; but the waves do not follow the great circle exactly. The true wave path is affected by the variation of velocity with depth, and must be determined in accordance with FERMAT's principle; but in the present case it cannot diverge very far from the great circle, as it crosses the major contour lines almost perpendicularly.

The period of the observed waves was about 15 minutes at both stations. This corresponds to a wave-length of nearly 200 km. The maximum height of these waves (from crest to trough) was about 15 cm. at Hilo, and about 12 cm. at Santa Barbara. As these records were obtained in shallow water, the amplitudes in the open ocean may have been appreciably less.

Since the wave-length is large compared to the depth of the ocean, the velocity V at a point where the depth is h is given by

$$V = \sqrt{g\,h}\,. \tag{16}$$

The travel time is then given by

$$t = \int \frac{d\,\Delta}{V} = \frac{1}{\sqrt{g}} \int \frac{d\,\Delta}{\sqrt{h}}\,.$$

The two travel times have been calculated by this method, the depths being read from the bathymetrical charts of the Pacific Ocean recently issued by the U. S. Hydrographic Office, Navy Department, Washington (H. O. Miscel. Nos. 8115 and 8461). It has been assumed that the waves reaching Hilo travel round by way of the south side of the island of Hawaii to the station (which is on the northeast coast). There must be a corresponding small deflection of the path to Santa Barbara, since the two stations lie nearly on a single great circle through the epicenter. The calculated travel times are 8.3^h and 13.4^h in good agreement with the observations.

VII. General summary on the structure of earth.

In this and the preceding papers seismic data of various types have been applied to the problem of the structure of the earth. It appears desirable to collect these scattered results and summarize them at this point.

There are two major discontinuities of the first order, which divide the earth into crust, mantle, and core.

The crust is itself divided in most regions into several layers; the uppermost is the layer of sedimentary rocks, with velocities of longitudinal waves from about 1 km./sec. in very unconsolidated recent material to at least 6 km./sec. in very old, consolidated sediments. Beneath these sedimentary rocks is a layer which in many cases is known to consist of granitic rock, in which the velocity of longitudinal waves is about 5.5 km./sec. In some regions the sediments are directly underlain by basaltic rock; in such regions the notation P_g, etc., should not be used. Where the sedimentary layer consists of very old, consolidated rocks, it is not impossible for the second layer to have a lower velocity than this first layer; for example, this may apply in parts of northern Germany or of the eastern United States, where high velocities have been found near the surface.

In several regions one or two deeper layers have been recognized within the crust, but their material has not been identified.

The thickness of the sedimentary layer varies locally within very wide limits; it may be totally absent, or may extend to depths of over 12 km. [Depths of this order have been found [37]) in the Los Angeles Basin by the use of applied seismic methods.]

The base of the granitic layer has been found, in the continental regions where it has been studied, at depths between 15 and 20 km. In these same regions the total thickness of the crust (depth of the first major discontinuity) has been found to be from 30 to 50 km. Relatively small values for this thickness have been found for the southwestern United States, western Europe, and northeastern Japan; about average thicknesses occur in central and western North America, and in South America. The largest values found thus far are in the region of the Alps. In the Atlantic and Indian Oceans the total thickness of the crust is only a fraction of that on the continents; the seismological data offer no evidence as to the nature of the rocks composing the crust in these areas, but in both oceans there still is a well-marked discontinuity between the crustal rocks and the mantle. There is no evident vertical discontinuity between these oceans and the adjacent continents.

In the region of the Pacific basin no marked discontinuity between crust and mantle exists; except for local accumulations of erupted basaltic material, it does not appear that the elastic constants near the rock surface differ significantly from those in the mantle. Data for the north polar basin definitely indicate the existence of a considerable area with properties similar to those of the Pacific basin.

All available evidence indicates that a continental type of structure exists in certain outlying areas of the Pacific Ocean. This is the case in the Polynesian region, including the area west of the Bonin, Marianne, and Caroline Islands. As mentioned in our second paper [2]) (page 315), there is evidence for continental structure in a limited area in the southeastern Pacific, at considerable distance from the coast of South America.

In the outer part of the mantle the velocity (at least for longitudinal waves) is nearly constant, and is the same for all regions (about 8.0 km./sec.). It is even possible that this velocity decreases slightly with depth near a level 70 to 80 km. below the surface of the earth. At about 100 km. it begins to increase gradually with depth; we find no evidence of any first-order discontinuity within the mantle, but a discontinuity of the second order appears to exist at about 1000 km., and two others near 2000 km. The first of these is that originally discussed by WIECHERT[30]); he assigned it a depth of 1521 km., which he considered as the boundary between the mantle and core. The more accurate data now available result in a smaller depth, while the core turns out to be at about 2900 km., where there is a discontinuity of the first order.

The hypothesis of a two-layered earth (mantle and core) was advanced by WIECHERT in 1897[31]) in order to account for the observations on density, the flattening of the earth, distribution of gravity and the movement of the axis of the earth. The first to find evidence of the existence of a core from seismic data was OLDHAM who in 1906 published a paper using the data of the Indian earthquake[32]). He found a depth of 0.6 of the radius (roughly 3800 km.); this result is in part due to combining observations of P, PP, and P' into a single supposedly continuous curve, and using for the S curve observations of S, PS, and SS. The first determination on correct principles was made in 1912[33]).

We find no unimpeachable evidence for the existence of transverse waves in the core. Due to the complexity of the seismograms involved, it is not possible to assert positively that no such waves exist. Calculations based on the data on the rigidity of the earth as a whole[34]) support the hypothesis of a core with small or zero rigidity (liquid). We cannot be certain of the physical meaning of this result; it is highly doubtful whether it can be interpreted in such simple terms as "crystalline" mantle and "molten" core. The deeper parts of the mantle may well be in an amorphous condition, but still solid; in any case it is doubtful how far these terms apply to matter under such extreme conditions.

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